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Problems Related to the International (Standard) Stratigraphic Scale and Its Perfection

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Abstract—All the aspects of the International stratigraphic scale (ISS) versions prepared by the International Commission on Stratigraphy in 2000 and 2002 are analyzed to show their similarity with and differences from the General stratigraphic scale (GSC) accepted and used in Russia. It is shown that the Precambrian ISS based on chronometric data is unacceptable, because geological systems of the Precambrian, except for the Vendian, cannot be defined. As is substantiated, the boundary and ISS stratotypes are complimentary, and the former cannot substitute the latter. The idea of validating the Phanerozoic ISS units depending on approval or rejection of the Global Boundary Stratotype Section and Point (GSSP) is criticized. It is shown that placing the Holocene and Pleistocene into the Neogene System, as it is done in the ISC-2002, we come to elimination of the autonomous Quaternary System. The attitude of the Interdepartmental Stratigraphic Committee of Russia to the procedure of consideration and acceptation of important international resolutions concerning general stratigraphic aspects is formulated. Substantiation of stratigraphic boundaries for stages established in Russia is supposed to be a procedure aimed at ISS perfection.

Key words: International (Standard) stratigraphic scales of Precambrian and Phanerozoic, unit and boundary stratotypes, Global Boundary Stratotype Section and Point, Quaternary System.

The perfection of the International (standard) stratigraphic scale, the elaboration level of which reflects achievements in stratigraphy, is among the most important tasks of this science.

By the middle of 20th century, the ISS represented a sufficiently stable structure with succession and ranks of its subdivisions accepted worldwide, and its perfection progressed toward specification of stage boundaries and further detailing (definition of substages, zonal and infrazonal biostratigraphic units). At that time, ranks of some systems (Silurian, Permian) were under discussion, and new stages were introduced into some systems (Cambrian, Silurian, Devonian, Permian, Neogene). The specified position of stage boundaries precisely was regarded to be critical for reliable regional, interregional, and global stratigraphic correlations.

A long-term verification of the Silurian–Devonian boundary by the special committee of the International Commission on Stratigraphy (ICS) in 1960–1972 under the leadership of H.K. Erben (Germany) and then of A. Martinsson (Sweden) is a perfect example methodically correct work. It should be mentioned that specialists from 21 countries took part in these study and in field examination of boundary sections in Belgium, Germany, Czech Republic (Barrandien), Morocco, the United States (Nevada), and the USSR (Podolia, Salair). A serious book summarizing the study results includes descriptions of over 20 sections and brief characteristics of 13 faunal and floral groups. This was a real scientific substantiation of that boundary, which resulted in recognition of new Pridolian, Lochkovian, and Pragian stages of the systems (*The Silurian–Devonian Boundary*, 1977).

Of course, it is hardly possible now to substantiate all the debatable boundaries between the ISS units based on similar long-term and expensive researches. Nevertheless, the methodological approach should remain precisely the same, if we want to achieve desirable.

Unfortunately, further studies developed in line with a simplified scenario turned finely into selection and approval of the Global Boundary Stratotype Section and Point (GSSP).

In the 1980s, leaders of the International Commission on Stratigraphy commenced actions aimed at perfection of the International stratigraphic scale that was called global. In 1986, J. Cowie, W. Ziegler, A. Boucot, M. Basset, and J. Remane proposed a project of the Global stratigraphic scale with a stage as its lowest-rank unit for a wide discussion (Cowie *et al.*, 1986; Cowie, 1986). Their project included the rules of the GSSP selection and description (*Stratigraphic Code*, 1992, Appendix 2). Some comments of the Interdepartmental Stratigraphic Committee of the USSR to the Global scale were forwarded to the ICS.

At the 28th Session of the IGC in 1989, Cowie and Basset (1989) presented, as editors, a new version of

the Global stratigraphic scale added by geochronological and magnetostratigraphic data. In principle, the scale remained the same, although it was slightly modified.

Based on materials presented by all the commissions on systems and subdivisions of the Precambrian, the Stratigraphic Committee of Russia prepared and sent to the ICS circumstantial letters (of February, 1990, and of March, 1992) with comments to coincidences and differences between the ISS and GSS accepted in Russia. In his letter of May 12, 1993, J. Cowie informed that the GSS accepted in Russia would be most useful for the ISS perfection.

J. Remane (Swiss), the new chairperson of the ICS elected at the 29th Session of the IGC continued the work of his predecessor. In 1998, the preliminary version of the International Stratigraphic Scale was published. It was approved by the Commission on the Geological Map of the World (CGMW) and by the International Union of Geological Sciences (IUGS) as it is evident from accompanying emblems (*International Stratigraphic Chart*, 1998).[1] J. Remane declared that the preliminary edition was prepared in cooperation with all subcommissions of the ICS and that the editorial working group included J. Dercourt and P. Bouysse, leaders of the CGMW, and A. Faure-Muret who took part in compiling the Geological Atlas of the World. Simultaneously, V.E. Khain, Yu.G. Leonov, and G.E. Grikurov, members of the CGMW bureau, reported that opinion of the CGMW plenary meeting was not unanimous at all. The Resolution of 1998 included suggestion to take into consideration the results of discussion at the General Assembly of the CGMW in further edition of the scale before its presentation at the 31st Session of the IGC in 2000.

The scale of 1998 met numerous objections of Russian stratigraphers. A new circumstantial letter, which summarized opinions of the ISC subcomissions, was sent after its approval at the enlarged ISC bureau, to J. Remane, A. Salvador (former chairperson of the International Subcommission on Stratigraphic Classification, ISSC), and to A. Riccardi, the editor of the second edition of the International Stratigraphic Guide and acting chairperson of the ISSC. The last two addressees informed that they do not also agree with some innovations in the scale. The letter was published in the "Resolutions of the Interdepartmental Committee" of 1999 and in official ISSC Circular (1999). The letter specially emphasized that the project of the International stratigraphic scale needed in discussion by national stratigraphic bodies. As is clear now, J. Remane was not going to follow this suggestion and presented at the 31st Session of the IGC (Rio de Janeiro, 2000) the International stratigraphic chart (colored and with explanatory notes) similar in principle to the version of 1998. Moreover, the presented version included some unsubstantiated renovations. The emblems of UNESCO and IUGS meant to confirm the status of the chart (*International Stratigraphic...*, 2000).

The International Stratigraphic Chart of 2000 (ISC-2000) was considered briefly by all the ISC commissions on Precambrian systems and subdivisions and twice at meetings of the ISC bureau in September of 2000 together with members of the Scientific–Editorial Council of the Ministry of Natural Resources and at its enlarged meeting in February of 2001with myself as a speaker (Zhamoida, 2003). After discussion of the International Stratigraphic Chart and received comments to it, the ISC Bureau approved the detailed resolution, which was forwarded to the Scientific–Editorial Council of the MNR. The resolution emphasized a significant similarity between the International Stratigraphic Chart and the General Stratigraphic Scale accepted in Russia as a statutory standard (*Stratigraphic Code*, 1992). Features in common were as follows: subdivision of the scale into three largest units, the Archean, Proterozoic, and Phanerozoic (they are termed as eonothems and acrothems in the ISC and Russian GSS, respectively) with close ages of lower boundaries accepted for two upper units; traditional systems, series (except for the Silurian, Carboniferous, and Permian), and stages (except for the Cambrian, Ordovician, and Upper Permian) accepted in the scale; similar letter indices for erathems, systems, and series (except for the Cambrian and Paleogene) and identical coloration of most subdivisions. For the first time, the official ISS includes the autonomous Paleogene and Neogene systems, Induan and Olenekian stages of the Triassic (both systems and stages used in the USSR since 1956 were approved by the ISC resolution) and all "Russian" Carboniferous and Lower Permian stages. At the same time, the International Stratigraphic Chart of 2000 contains several renovations, which do not appear to be sufficiently substantiated. The approval of some renovations can result in serious violations of traditional rules in geological mapping used by most national geological surveys and in international projects.

The resolution on the ISC-2000 was published in *Resolutions of the Interdepartmental Committee* of 2002 and in regular ISSC Circular (2001) and sent to the International Commission on Stratigraphy and to M.B. Cita (Italy), the current chairperson of the ISSC. Some Russian geologists accepted showy copies of the ISC-2000 accompanied by explanatory notes as the approved official international document. Therefore, several Russian journals published articles, where the statute of this chart was clarified and which included relevant resolutions of the Russian ISC (Zhamoida, 2000; Zhamoida and Prozorovskaya, 2001; Gladenkov, 2001).

Gladenkov (2002) represented officially the Russian ISC at the working meeting of the International Commission on Stratigraphy (Urbino, Italy, 2002) dedicated

[1] In Russian, the English word "chart" corresponds in the considered case to word "scale" (*Addendum to the Stratigraphic...*, 2000).

to the present problems and prospects of stratigraphy. He informed about reports of all chairpersons of subcommissions on systems (from the Cambrian to Quaternary) on the present-day knowledge of various systems and on plans for 2004–2006. It was decided at that meeting to prepare and distribute a new version of the International Stratigraphic Scale with recent data on subdivision and dating of ancient sections by 2004, i.e., by the 32nd Session of the IGC.

The results of the ICS working meeting in Urbino are briefly presented in the article by F. Gradstein, the chairperson of the Commission, and J. Ogg, its general secretary (Gradstein and Ogg, 2002). The article is supplemented with the chart of the Overview of GSSP (compiled by Ogg) lacking letter designations of the International stratigraphic scale.[2] The Phanerozoic stages are accompanied by a brief and very incomplete characterization of their lower boundaries, sometimes with indication of the Principal correlative events, by the name and location of the GSSP, boundary status (approved or debatable), and reference publication (mostly from journal *Episodes*). Magnetostratigraphic data are shown for the lower boundaries of the Neogene stages and the Aptian, and the Ir-anomaly is shown at the Danian Stage base.

The principal difference of the Phanerozoic stratigraphic scale under consideration (ISC–2002) from previous versions consists in elimination of the Quaternary System as an autonomous system (period), because the Holocene and Pleistocene are included into the Neogene System.

PRECAMBRIAN

The Precambrian portion of the International stratigraphic scale of 2002 is identical to that in the ISC-2000: boundaries of subdivisions are characterized here by the same approved Global Standards of Stratigraphic Ages (GSSA). In its structure, nomenclature of subdivisions, and boundary ages, the Proterozoic part of the new scale corresponds to the scale approved by the International Subcommission on Precambrian Stratigraphy (Plumb, 1991). Therefore, the former assessment of the Precambrian scale formulated in the resolution of the ISC Bureau of February 2, 2001 (*Resolutions of the Interdepartmental...*, 2002) remains valid and consists in the following:

(1) The ISC cannot accept the principle of the Precambrian scale constructing based chronometric data only, i.e., the refusal from the uniform approach to both the Precambrian and Phanerozoic, which is of essential importance. By the way, Professor H.D. Hedberg, the editor and, in fact, the author of the first International Stratigraphic Guide (1976) was an advocate of this concept.

(2) The general Precambrian stratigraphic scale elaborated by Russian stratigraphers and based on the complex study of type sections with recognition of specific geological formations and analysis of isotopic dates obtained by different methods demonstrated a high efficiency of the historical–geological approach to subdivision and correlation of Precambrian sequences. Therefore, the ISC recommends to use the Lower Precambrian stratigraphic scale approved at the 3rd All-Russian meeting (Apatity, 2000) and Upper Precambrian (Upper Proterozoic) scale approved at the 2nd All-Union meeting (Ufa, 1990) with specified isotopic dates for the Upper Riphean, Lower and Upper Vendian, and Cambrian lower boundaries published in *Addenda to the Stratigraphic Code of Russia* (2000). Both these scales were approved at enlarged meetings of the ISC Bureau and published (*Resolutions of the Interdepartmental...*, 1992, 2002; *Resolution of the 3rd All-Russian Meeting...* 2001).

(3) Recognition of geological systems in the Proterozoic is unacceptable except for the Vendian System that is comparable with the Neoproterozoic-III in the International Stratigraphic Chart. The stratotype of the Vendian System is comprehensively studied in the East European platform and its reference section in carbonate facies, in the Siberian platform.

According to recent data, the isotopic age of the Vendian and Upper Vendian lower boundaries is accepted to be at 600 ± 10 Ma and within the interval of 570–555 Ma, respectively (*Addenda to the Stratigraphic Code...*, 2000). Stratigraphic analogues of the Vendian are established in southern China, Australia, South Africa, Canada, and in the British Islands. The Vendian System is included in serial legends of State geological maps of Russia and widely used by geological mapping in Ukraine and Belarus (*The Vendian System*, 1985; Sokolov, 1997).

UNIT AND BOUNDARY STRATOTYPES

Before going to analysis of the general Phanerozoic stratigraphic scale, we should dwell on the problem of relationship between a stratotype of subdivision in the general scale and the lower boundary stratotype of that subdivision (or limitotype). This issue was in detail considered in the report to the International Symposium "Upper Permian Stratotypes of the Volga Region" (Zhamoida, 1999), therefore I repeat here only some principal points.

In 1996, the ICS approved the statute of the Global Boundary Stratotype Section and Point (GSSP). Although the attitude of geologists to this concept is different (Naidin, 1998; Zhamoida, 1999; Gladenkov, 2001; and others), the proposed procedure of selecting the lower stage boundaries is approved by international stratigraphic bodies and consequently, is obligatory. Description of the procedure is given in the Addendum 2 to the Russian Stratigraphic Code of 1992 and in

[2] Let us conditionally term it as the ISC-2002.

Table 1. Amount of Phanerozoic stages in the International Stratigraphic Charts of 2000 (2002) and their status according to the International Commission on Stratigraphy (Table 1, Gradstein and Ogg, 2002)

GSSP / Geological systems	Number of stages in ISC-2000 and ISC-2002	Number of approved GSSP, years of approval			Candidates for approval	Guide events at boundaries not defined
		ISC-2000	Added to ISC-2002	Total number		
Cenozoic	18	8 / 1985–97	–	8	3	5
Mesozoic	30	2 / 1996–2000	4 / 2000–02	6	9	8
Carboniferous–Permian	16	3 / 1990–96	3 / 2001	6	–	4 (Carboniferous)
Silurian–Devonian	15	15 / 1972, 1980–95	–	15	–	–

Addenda to the Stratigraphic Code of Russia (2000). At the same time, it is impossible to agree that the stage validity depends on the approved GSSP. According to this approach, almost all Mesozoic stages (except for the Aalenian and Bajocian) would be unofficial in the ISC-2000, although they are recognizable in practice more reliably than some Paleozoic stages with approved GSSPs. Moreover, despite urgent requests of the ISC, substantiated GSSPs for different intervals of the general scale are proposed very irregularly (Table 1), and this is likely not incidental.

The first International Stratigraphic Guide (1976) points to significance of the stratotype for chronostratigraphic (general) units. In the second edition of the International Stratigraphic Guide (1994), only the boundary stratotypes, i.e., the GSSPs are considered obligatory for general subdivisions. Nevertheless, A. Salvador, its editor considered the study of unit stratotypes desirable at least, understanding a universal significance of this procedure (ibid, p. 88).

Authors of the *Revised Guidelines for the Establishment of Global Chronostratigraphic Standards by the International Commission on Stratigraphy* (Remane *et al.*, 1996) are of a principally different opinion. In the sections entitled "Boundary-stratotypes instead of unit-stratotypes," they unequivocally state: "Chronostratigraphic units of the Phanerozoic Global Standard can only be defined through boundary stratotypes" and further: "The use of terms like holostratotype, parastratotype, etc. should therefore be avoided" (ibid, p. 78). Thus, according to the International Stratigraphic Guide (1994) and the guidelines mentioned above, the unit stratotypes of the General stratigraphic scale will turn into an exception in the geological practice.

Gladenkov (2001, p. 111) justly notes that the passion for "points" and "nails" in stratigraphy reduces to zero the analysis of the ISS stratigraphic units proper, actually belittling and even excluding the regional stratigraphy out of the "ISS perfection process." Meanwhile, the International (General) stratigraphic scale was itself constructed based on regional stratigraphy or, as was noted by Sokolov (1971, p. 160) is a "consequence of the purely regional synthesis."

Another important point is as follows. To select and substantiate in line with the accepted procedure, and to approve the GSSP is only half the work. The main objective is to establish this boundary in particular sections worldwide. Moreover, sediments corresponding to various units of the General stratigraphic scale should be correlated.

The International Stratigraphic Guide (1994) and leaders of the ICS (Remane *et al.*, 1996) recommend a wide application of non-biostratigraphic methods for GSSP selecting. The suggestion is correct. Nevertheless, despite doubtless achievements, the isotopic, magnetostratigraphic, geochemical, paleogeographic, and other methods are far from being self-sufficient for remote correlation. The recommendation to apply the sequence stratigraphy method when selecting the GSSP looks even stranger. Authors of the Guidelines themselves consider the biostratigraphic method as the main one for definition of boundaries between Phanerozoic units.

A stage, the stratotype of which materializes the corresponding interval of the scale, is the main stratigraphic unit in the General stratigraphic scale of the Phanerozoic. Boundaries between series and systems correspond to the stage boundaries. Specification of boundaries between stages is specification of boundaries between zones. The first question is therefore: which interval of the section should be included into the boundary stratotype? Is it enough to take one zone from each conjugate unit or should we take two–three zones from them for the reliability? What should we do if there are only two or three zones in the stage? In this

case, practically the entire stage, i.e., its stratotype, should be taken into account.

Another question concerns the so-called conjugation of successive zones. For the Phanerozoic, only successive changes in the fossil faunal (floral) assemblages (following the Hucksley's principle) or in continuous phylogenetic lineages enable a relative objectivity in selecting the events, which are important for recognition of any boundary stratotype, the GSSP inclusive. This approach precisely is recommended for the GSSP substantiation, although it is incompletely reasonable, because only two levels (base and top) but not the whole section (unit stratotype) are under consideration.

"Extracting" from the section the boundary layers only by selecting the GSSP, we encounter the other uncertainties. First, significance of several plant groups (macroflora, spore-pollen assemblages, charophytes) recedes, because boundaries of relevant units are frequently established based on changes in quantitative distribution of the same taxa (domination degree). The same is true of some freshwater fishes and benthic animals. Given these fossil groups useless, it will be naturally difficult to establish boundaries between the GSS units in continental sections. It should also be noted that recommendation to use solely sections composed of pelagic sediments when selecting the GSSP is disputable, particularly for the systems represented mostly by continental sediments, volcanogenic facies included. In general, using as many fossil groups as possible by characterization of any stratotypes (boundary stratotypes included), we get a highest efficiency of stratigraphic correlations.

The idea to select a narrow interval for the boundary does not account for some phenomena such as, for instance, the appearance of species and genera at different stratigraphic levels in different regions. This phenomenon improperly stipulated is responsible for the notion of "transitional beds" and "time belt" in the Norwegian stratigraphic code (*Regler for Norsk Stratigrafisk Nomenclatur,* 1961).

When the stratotype is ignored, difficulties and even impossibility are unavoidable in tracing boundaries between the GSS units because of "sedimentological factors."

In condensed or lithologically uniform sections, boundaries between units, for example between stages, can be interpolated from more distinct marker horizons located below or above, i.e., beyond the boundary stratotype interval. In sections of most systems, there are the correlative levels of this kind recognizable in different continents. The same approach should be used in case of hiatuses, event though they are short, either within sediments of some GSS unit or, in particular, in the boundary intervals. Even if a boundary is reliably established in areas located far away from the boundary stratotype, it is always useful to define marker horizons (biostratigraphic ones included) to enhance validity of that boundary.

Neglecting the unit stratotype, we practically exclude applicability of rhythmostratigraphic methods (sensu lato), which are useful for intercontinental correlation, although to a lesser extent than some researchers believe. The same is true of the so-called event-stratigraphy method very popular now. In fact, stratigraphy since its appearance is completely based on registration of events, although mission of event-stratigraphy is to define those boundary events recorded in a section, which were much shorter than events recorded below and above in the succession. It is clear that this is hardly attainable in a narrow boundary interval desirable for the GSSP. Thus, the unit stratotype disavowal appears to be a great error as far at it concerns the Phanerozoic subdivisions of the general stratigraphic scale. The boundary stratotype is not an alternative to the unit stratotype of any kind, the GSS stratigraphic subdivisions included.

The unit and boundary stratotype are complementary notions. It seems that adherents of the unit stratotype denial are interested in the establishment and official approval of the GSSP rather than in possibility of practical tracing of given boundaries in particular intricate situations.

PHANEROZOIC

I am in bewilderment why all the Cambrian stages became excluded from stratigraphic scales of 2000 and 2002, especially the "Russian" stages of the Lower and Middle Cambrian, which were included into the Global stratigraphic scale of 1989 (Cowie and Busset, 1989). In Russia, geologists use the scale of the Cambrian System recommended by the Stratigraphic Code (1992) with addition of the Upper Cambrian Batyrbaian Stage of 1997, which overlies the Aksaian Stage (*Resolutions of the Interdepartmental...*, 1997). The isotopic age of the Cambrian lower boundary in the ISC-2000, ISC-2002, and in the Russian scale is estimated to be 540 ± 5, 543.3, and 535 ± 1 Ma, respectively (*Addenda to the Stratigraphic...*, 2000).

In the scale of 2002, boundaries of the third (Middle Ordovician) and fifth (Upper Ordovician) yet unnamed stages are approved, in addition to the lower boundaries of the Tremadocian and Derrivilian stages officially approved for the ISC-2000. Such a situation and still "uncertain" status of the Derrivilian Stage are hardly admissible from the standpoint of the general stratigraphic scale.

Russian stratigraphers cannot accept subdivision of the Silurian System into four series as proposed in the ISC-2000, because this act violates traditions. The Silurian is subdivided in two series: the lower one including the Llandoveri and Wenlock and upper series consisting of the Ludlow and Přidoli. The proposed units termed as stages in the ISC-2000 are used in Russia in the substage rank.

All the stages of Devonian System have their lower boundaries approved; the Silurian–Devonian boundary was approved in 1972.

In the ISC-2000 and ISC-2002, the Carboniferous System is subdivided in two, the Mississippian and Pennsylvanian subsystems. The former corresponds to the traditional Lower Carboniferous (with Tournaisian, Visean, and Serpukhovian stages) and the latter can be subdivided in two series with ranges determined by stages of the East European scale: the lower series (middle series of the system) of the Bashkirian and Moscovian stages and the upper series of the Kasimovian and Gzhelian stages. In opinion of the ISC Commission on the Carboniferous System, the Tournaisian and Visean stages can be regarded in the rank of superstages divisible into stages (two in each) corresponding to well-substantiated regional units of the East European scale.

In the ISC-2002, only lower boundaries of the system and Pennsylvanian Subsystem (Bashkirian Stage) are considered as officially approved. As for the boundaries of four remaining "Russian" stages, it is stated there: "Guide event is undecided." The wording should alarm Russian stratigraphers who must undertake resolute actions for circumstantial argumentation to substantiate validity of these stages. This means the accomplishment of formal, though rather strict requirements for the GSSP approval.

The Permian System of the ISC-2000 is proposed to be divided into three series. It should be noted that Russian stratigraphers repeatedly discussed a possibility of three-member subdivision as well. Names of the series (Cisuralian, Guadalupian, and Lopingian) violate traditional rules of the stratigraphic nomenclature, although they seem to be justified as eliminating the alternative reading. The lower system boundary only is approved in the ISC-2000, and thus, only the Asselian Stage is admitted. In 2001, the GSSPs of all three stages of Guadalupian series and, consequently, its basal boundary were approved. The International Subcommission on Permian Stratigraphy (ISPS) presented the lower boundary of the Lopingian Series for the approval. The Russian experts on Permian stratigraphy undertake intense actions to substantiate the remaining three "Russian" stages of the lower Cisuralian Series and to present them for the official approval. This issue is to be discussed at the General Symposium G-22.04 (32nd Session of the IGC) "Global Correlation of Cisuralian stages (Lower Permian)" with B.I. Chuvashov as one of conveners. Inasmuch as the Permian Period was a time of widespread continental sedimentation, the Interdepartmental Stratigraphic Committee of Russia suggests to approve officially two parallel scales for the Upper Permian (Guadalupian + Lopingian): the East European scale for the Biarmian and Notal realms and for other regions with continental Permian sediments, and the ISS version for the equatorial and Tethyan regions. The latter cannot practically be used in the other realms and regions.

This proposal was supported by participants of the International Symposium "Upper Permian Stratotypes of the Volga Region" (*Reports...*, 1999), Prof. B. Wardlow, the chairperson of the ISPS included. The scale proposed by the Interdepartmental Stratigraphic Committee can be accepted as the basic (main) one.

The recent finds of Roadian ammonoid and conodont assemblages in basal strata of the Kazanian Stage and Wordian conodonts found in Kazanian deposits of various regions of Russia (Novaya Zemlya, Verkhoyan'e, southern Primor'e, and others) offer an opportunity of direct correlation between the East European scale and ISC-2000.

All the Cretaceous, Jurassic, and Triassic stages of the international stratigraphic scale were accepted in Russia as official units and were always used in geological legends of various kind and in stratigraphic studies. Nowadays, they are in use as formerly in Russia and in all northern countries of the former USSR. In opinion of the Interdepartmental Stratigraphic Committee, it would be unreasonable and inadmissible to rename or replace the Mesozoic stages. As for the Mesozoic of the ISS, the Induan (Triassic lower boundary), Sinemurian, Aalenian, Bajocian, Cenomanian, and Maastrichtian stages are considered in it as officially approved.

The lower boundaries of the Paleogene (Danian Stage) and Neogene (Aquitainian Stage) systems are officially not approved. The Paleogene–Neogene scale used in Russia is identical to that of the ISC-2000.

QUATERNARY SYSTEM

As is known, the Quaternary System of the Pleistocene and Holocene is broadly acknowledged, being in use since the first quarter of the 19th century. It is included in all international and regional (national) stratigraphic codes. The International Union for Quaternary Research (INQUA) with its Commission on Quaternary Stratigraphy exists since the 1930s.

In the former USSR, the Commission on Quaternary Research under the aegis of Academy of Sciences was formed in 1927 on initiative of academician V.I. Vernadskii. It was successively headed by academicians A.P. Pavlov, F.Yu. Levinson-Lessing, I.M. Gubkin, V.A. Obruchev, V.N. Sukachev, G.I. Garetskii, and A.L. Yanshin. Since organization of the Interdepartmental Stratigraphic Committee in 1955, the commission on the Quaternary System actively collaborated with this institution. In Russia, the Pleistocene was traditionally divided into the lower and upper subseries.

In 1995, the subseries were named by the ISC approval as Eopleistocene (corresponds to the lower Pleistocene or Calabrian of the European scale) and Neopleistocene (corresponds to the middle–upper Pleistocene of the European scale) (Table 2, *Resolu-*

Table 2. General stratigraphic scale of the Quaternary System (*Addenda to the Stratigraphic Code of Russia*, 1992, p. 23)

General stratigraphic units					Principal geochronologic boundaries (Ma)	Geochronologic units				
system	series	subseries	link	step		period	epoch	phase	time	thermochron, cryochron
Quaternary	Holocene				–0.01–	Quaternary	Holocene			
	Pleistocene	Neopleistocene	upper	forth			Pleistocene	Neopleistocene	late	late cryochron
				third						late thermochron
				second						early cryochron
				first						early thermochron
			middle						middle	
			lower						early	
		Eopleistocene	upper					Eopleistocene	late	
			lower		–1.8–				early	
Neogene	Pliocene	Upper				Neogene	Pliocene			

tions of the Interdepartmental..., 1996; *Addenda to the Stratigraphic Code...*, 2000).

The Quaternary, as an autonomous system consisting of two series (Holocene, Pleistocene), was included into the ISC-2000 and discussed by Gradstein (2000). Nevertheless, in their article on scientific perspectives of stratigraphy, Gradstein and Ogg (2002) eliminated the Quaternary System from the chart and included Holocene and Pleistocene into the Neogene System without any reasoning. Their chart was reproduced in the ISSC Circular (2002) also without necessary comments.

The chart includes principal correlative events marking lower boundaries of the Holocene and three Pleistocene subseries. The Pleistocene lower boundary is dated at 1.806 Ma (the base of the Vrica sapropel beds in Italy) and the Holocene lower boundary at 0.1 Ma, i.e., as in Russia. There is a peculiar comment to the Neogene System: "The "Quaternary" (quotation marks as in original, A. Zh.) is traditionally considered to be the interval of oscillating climatic extremes (glacial and interglacial episodes) encompassing Holocene and "Pleistocene" (Gradstein and Ogg, 2002, p. 204).

The Interdepartmental Stratigraphic Committee of Russia and Commission on the Quaternary Research of the Russian Academy of Sciences could not agree with such the factual abolishment of the Quaternary as an autonomous geological system of the Cenozoic. The Commission on the Quaternary System of the ISC (S.M. Shik, B.A. Borisov, A.N. Alekseev, Yu.A. Lavrushin, E.P. Zarrina, V.D. Tarnogradskii) and bureau of the committee prepared a relevant letter signed by myself, which was forwarded to the ICS, ISSC, INQUA, and its Commission on the Quaternary Stratigraphy (*Resolutions of the Interdepartmental...*, 2003).

The long-term studies in Russia and other countries brought out clearly that in its geological history the Quaternary sharply differs from the Neogene and from older periods. Characteristic of the Quaternary period were recurrent continental glaciation in middle latitudes (these events have nor been recorded here since the Paleozoic) and alternation of pluvial and arid epochs in low latitudes. These peculiar features of the Quaternary determine a wide distribution of glacial sediments and loess-and-soil formations. These sediments and formations of a great practical significance

for ecology and engineering geology require application of specific methods by the study and geological survey. Quaternary sediments up to several hundreds meters thick are widespread in sedimentary cover of seas and ocean floor and cover the surface of continents. There are two kinds of geological maps used worldwide: of the Quaternary and pre-Quaternary sediments. Maps of the first kind usually depict data on age, genesis, and lithology of sediments, as it is important for practice.

The periodicity of glaciations offers an opportunity to apply climatostratigraphic methods for subdivision and correlation of Quaternary sediments. In combination with biostratigraphic methods (small mammals, landscape palynology, and others), this approach enables definition and mapping of subdivisions corresponding to very short time spans (around 100 k.y.) that is impossible for older, even Neogene sediments.

The appearance and development of the genus *Homo* is a unique feature of Quaternary period that stimulated development of archeological stratigraphic methods unsuitable for older periods of the Earth history. That is why Russian academician A.P. Pavlov proposed in 1919 the term "Anthropogene" for the Quaternary System.

The Quaternary System crowning the General (International) stratigraphic scale differs from others Phanerozoic systems so significantly that it turned into mission of the Quaternary geology, an autonomous complex scientific branch dealing with peculiar features of Quaternary sediments and operating with specific methods of their study. Among specific problems it is addressed to ecological (sensu lato) aspects of stratigraphy are most important. The program of the 32nd Session of the IGC includes General Symposium G 19: Quaternary Geology.

In conclusion, the letter of the ISC emphasizes the validity of the Quaternary System consisting of the Pleistocene and Holocene and attracts the ICS attention to potential difficulties we can meet by geological mapping and by comparison of new and former geological maps and other materials because of changes in the international stratigraphic scale.

In July 2003, I have received the reassuring letter (of July 1, 2003) from F. Gradstein, the chairperson of the ICS, in which he said: "As is usual in Stratigraphic Science, there is a distinction between formally approved and informal stratigraphic names. The ISC fully respects the use of the informal name Quaternary, and it is included in the official time scale at the official website of ICS. An exhaustive documentation of the history and current use of various Cenozoic terms will appear in the forthcoming book "The Geologic Time Scale" (Cambridge University Press) co-authored by 40 stratigraphers. The formal program of ICS does not suppress the name Quaternary. ICS hopes to be able to ratify the definitions of the Middle and Upper Pleistocene Boundaries, as well as the Holocene in the coming years.

It is unknown so far, what an "exhaustive documentation" will appear in a new book *The Geologic Time Scale*, but M.B. Cita, the chairperson of the ISSC considers the Gradstein's position as erroneous, and this may have unexpected consequences (her letter of July 8, 2003). She informed that INQUA is going to organize a new Commission on Stratigraphy, independent on the International Union of Geological Sciences and subordinate ISC.

GENERAL INFERENCES

(1) It should be once more emphasized that universal documents like the International (General) stratigraphic scale and their changes must be discussed by national stratigraphic bodies, not only by leaders and by several members of ICS subcommissions.

(2) The ISC should officially determine, in coordination with the International Commission on the Geological Map of the World, the status of the proposed International stratigraphic scale, i.e., clearly indicate for what kind of geological works it is obligatory and for what kind is recommended only.

In my opinion, the ISS can be obligatory for all international geological projects, which are financially supported and headed by international scientific bodies. Inasmuch as national geological traditions substantiated by the practice are strong in some countries, the International stratigraphic scale can only be recommended for national or international projects, if international geological bodies do not participate in the latter.

(3) Subdivisions of the ISS (GSS) should have approved the unit stratotypes and the GSSPs, which should be considered as supplementing characteristics for each unit, not alternative notions. The phrase "Boundary stratotypes instead of unit stratotypes" should be excluded from the "Revised Guidelines" to establish the Global Boundary Stratotype Section and Point. The validity (official status) of ISS units cannot be considered as justified only after the approval of GSSPs (official, semiofficial, unofficial). The information on the approval of GSSPs should be published in official circulars of the ISC and its subcommissions in the journal *Episodes* and in International stratigraphic scales issued separately.

(4) In the International Stratigraphic Guide (1994) and ISC-2000, a stage is the lowest-rank unit. In the Russian scale, the lowest-rank unit is a zone, while a step is the smallest unit of the Quaternary System. If the stage remains the smallest unit in the official International stratigraphic scale, it would be desirable to have supplementary biostratigraphic zonations for each system, which must be substantiated by advanced data and approved as standard units. At any rate, the International stratigraphic scale should be supplemented by a chart with indication of two adjoining zones, the

boundary between which corresponds to the boundary between corresponding stages, as it is done long ago for the Jurassic System (*Resolution of Plenary...*, 1963).

(5) There is no officially approved priority principle in stratigraphy, therefore national stratigraphic guides and researchers consider it obligatory to a varying extent and some of them ignore it completely. The priority principle, sufficiently strict and simultaneously permitting necessary divergences should probably be accepted in stratigraphy. Sections 11 and 12 of the Russian Stratigraphic Code (1992) dedicated to the validity and priority of stratigraphic subdivisions can probably be a basis for elaboration of the principle.

(6) The special paragraph in resolutions of the national ISC of February 2, 2001 (*Resolutions of the Interdepartmental...*, 2002) states: "The stratigraphic survey of Russia reserves the right to follow Russian geological traditions in stratigraphy and national geological–cartographic practice and to ignore the ISC-2000 in the intervals, terminology, and nomenclature, which are insufficiently substantiated and inconsistent with mentioned traditions and practice."

This important decision approved by the ISC Bureau was completely included into the resolution that was forwarded to F. Gradstein in December of 2001 and published in the ISSC Circular (2001).

(7) Russian delegates to the so-called General symposia at the 32nd Session of the IGC in Florence (Italy, 2004) should report to participants on the concepts and comments to the ISC-2000, which are outlined above. The ISC Bureau will prepare a special report to the ICS and ISSC meetings.

(8) Russian stratigraphers face the very serious tasks aimed at perfection of the International (General) stratigraphic scale. To solve the tasks, they should undertake all efforts to get the official international approval of stages first defined in Russia. These are all the Cambrian (the Batyrbaian Stage stratotype is in Kazakhstan) and Permian stages, the Serpukhovian, Bashkirian, Kasimovian, and Gzhelian stages of the Carboniferous System, and the Olenekian Stage of the Triassic System. The urgent character of actions is primarily dictated by the potential priority loss, when the stages originally established in Russia will be replaced within various intervals of the general scale by new stages recognizable in other countries. Scientific activity in this direction is fundamental and of practical significance, because it is necessary to keep stability of the General stratigraphic scale that is accepted in Russia as a basis for all geologic-cartographic and other works. The preparation of necessary documents can be performed only in the course of planned scientific–research works with participation of specialists from Russian Academy of Sciences, Ministry of Natural Resources, and higher school under guidance of corresponding commissions of the Interdepartmental Stratigraphic Committee.

The ISC commissions on the Permian and Carboniferous Systems already worked successfully in that direction. As was mentioned, the problem of boundary stratotypes for the Lower Permian stages will be discussed at the 32nd Session of the IGC. The Commission on Cambrian Stratigraphy activated its work: a special meeting headed by A.Yu. Rozanov was organized in Novosibirsk. The work of the Commission on Triassic Stratigraphy aimed at substantiation of the Olenekian Stage official status should also be intensified. Although the lower boundary of the Induan Stage (Triassic System) is approved by the ICS in 2001, it is established in China, not in the stratotype area of Pakistan (*Episodes*, 2001). Zakharov *et al.* (2002) proposed the Triassic section of the Abrek Bay (southern Primor'e) as a candidate for the global stratotype of the Olenekian Stage lower boundary. Judging from presented materials, this boundary is sufficiently well substantiated.

Now, it is difficult to foresee a possibility to organize the state program of studies mentioned above. Nevertheless, after discussion of problems at the 32nd Session of the IGC, the program should be organized in order to get desirable results and to prepare required materials to the next IGC session.

Reviewers Yu.B. Gladenkov and V.A. Zakharov

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Some Problems and Main Orientating Points of Stratigraphy in the Early XXI Century

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Abstract—Principal trends in development of present-day stratigraphy are analyzed. Most important orientating points of stratigraphy in upgrading the global stratigraphic scale and interpreting geological processes are discussed.

Key words: stratigraphy, stratigraphic scale, geological processes.

INTRODUCTION

Stratigraphic studies of the last years impel us to evaluate anew the state-of-the-art, problems, and orienting points of stratigraphy, an important branch of geoscience. We should reassess new approaches to stratigraphic classification, the International Stratigraphic Scale (ISS), and should discuss necessary trends of stratigraphic activity in future. The relevant issues were under discussion in last proceedings of International Geological Congress (IGC) and in documents of the International Commission on Stratigraphy (ICS). This work is devoted to analysis of these materials. Such an analysis appears to be useful prior to the 32nd IGC Session to be held in Italy in summer of 2004. At present, before the congress, the Interdepartmental Stratigraphic Committee (ISC) of Russia is liable to and should discuss topical problems of stratigraphy and formulate the concrete suggestions on important problems of stratigraphic activity.

The 31st IGC, the last one in the XX century, took place in 2000 in Rio de Janeiro (Brazil). During that congress, stratigraphy was given much attention to. Referring a reader to my review of stratigraphic problems touched upon in Brazil (Gladenkov, 2001), I would like to note that one, most important of them, was a new version of the ISS. In 1998, J. Remane, the ICS President, sent the last ISS version to all countries for discussing it in national stratigraphic institutions. In April 1999, the ISC of Russia mailed its comments and suggestions in reply. The scale version distributed in Brazil was in fact a replica of the ISS of 1998, in which many proposals of the Russian ISC and, probably, of other organizations are ignored (*Explanatory Note...*, 2000).

In short, innovations in the scale are as follows:

– *Precambrian.* The scale of abstract geological time with chronometric units devoid of stratotypes and corresponding to numbered subdivisions (with rounded age values accepted for their boundaries) is actually suggested. The "Riphean" and "Vendian" systems are missed. The Archean (with four erathems) and the Proterozoic (with three erathems and ten systems) are regarded as eonothems.

– The Paleogene and Neogene are considered as individual systems.

– New subsystems and series are authorized, two (Mississippian and Pennsylvanian) in the Carboniferous and three (Lopingian, Guadalupian, and Cisuralian) in the Permian.

– The Induan and Olenekian stages are legalized in the Triassic. The Asselian, Sakmarian, Artinskian, and Kungurian stages traditional of Russia are left untouched in the Lower Permian. The Middle Permian is divided into—Roadian, Wordian, and Capitanian stages of "American" origin, and Wuchiapingian and Changhsingian stages of China are included into the Upper Permian. The Gelasian Stage got the official status in the upper Neogene.

– The Cambrian system is now free of stages, the Ordovician and Silurian systems lost some of their former stages.

– Boundaries are dated based on materials of the ISC subcommissions and on the time scale by G.S. Odin (*Explanatory Note...*, 2000).

– The introduced new symbols for the Cambrian and for the Paleogene systems.

– Indices of stages in each system (lower-case first letter of the system name) are numbered from the base upwards; in the Neogene (N), for instance, stages are labeled as n1–n9.

– The coloration patterns for the ISS units (not always satisfactory) and the above numbering mode are copied from the *Geological Atlas of the World.*

It is noteworthy that a stage is the lowest-rank unit of the ISS. The ISC attention for the last years was

focused on boundaries between stages that had positive and negative aspects, in my opinion.

In 1986, the ICS adopted the concept of the Global Boundary Stratotype Section and Point (GSSP), according to which the lower boundary of the ISS unit should be determined by selecting a point in the approved boundary stratotype. The point simultaneously defines the upper boundary of underlying unit distinguished on the same basis. Accordingly, stratigraphic units can be regarded now as (1) official (with approved GSSP), (2) semiofficial (GSSPs are under consideration in respective working groups of the ICS), and (3) informal. According to the sate-of-the art of 2002, 27 of 39 boundaries are formal in the Paleozoic, 6 of 30 in the Mesozoic, and 8 of 20 in the Cenozoic. It is interesting to note that the Mesozoic used to be an example of erathem subdivided in detail is now divided, from the formal standpoint, worse than other erathems.

In opinion of some specialists who participated the congress, the ICS efforts to establish the GSSPs ("golden nails") lead away from vital stratigraphic problems. Instead of thorough substantiation of stratigraphic subdivisions, the ICS subcommissions concentrated main attention in recent years on the "nails" for their lower boundaries (as it happened in practice), thus raising this procedure to the rank of a concept. Many stratigraphers believe however that such a procedure viewed from side of other stratigraphic concepts is nothing else but a technology (know-how), which should be in use but is unable to replace methods of historic geology by subdivision of stratigraphic sequences. The point-or-nail mania reduces to zero the analysis of the ISS units important for stratigraphic reconstructions, thus actually disparaging and even putting away the regional stratigraphy from the ISS upgrading.

As discussions showed, the suggested ISS version met objections of many geologists, which, unfortunately, were not taken properly into account. Some stratigraphers were of opinion that the ISS revision is done hastily and formally that is inadmissible for basic documents.

A separate problem that was under attention at the congress in Brazil is the improvement of stratigraphic classification. From this viewpoint, it was important that the abridged version of International Stratigraphic Guide was published before the congress (*International...*, 1999). Later on, the guide was translated into Russian, and its positive and debatable aspects were assessed recently (Gladenkov, 2003). The new ICS Bureau was organized in Brazil with F. Gradstein (Norway) as Chairman, R. Lane (USA) as first Vice-Chairman, S. Finney (USA) as second Vice-Chairman, D. Ogg (USA) as Secretary General, J. Remane (Switzerland) as past-Chairman.

In June 2002, Gradstein organized a workshop of the ICS in Urbino (Italy). Short information about the event was published in the *Episodes* (Gradstein and Ogg, 2002). About twenty specialists (mainly chairmen of subcommissions) took part in the workshop. Two principal problems discussed were the current state of the ISS (first of all, the validity of the subdivision of systems into stages) and the future of stratigraphy.

All the chairmen of subcommissions (or their representatives) commented the first problem. The state-of-the-art is shortly as follows.

Terminal Precambrian (A. Knoll, the subcommission chairman missed the workshop); according to his report of 2001, the subcommission strove to determine methods of subdivision, criteria of correlation, and position of the GSSP to get a satisfactory nomenclature for the terminal system of the Precambrian. Subdivisions well substantiated in Russia, the Vendian included, were likely regarded as regional.

Cambrian (J. Shergold). The subcommission worked to elaborate a new stage scale (with new units and new names) with due regard to selected GSSP. For Russian geologists, such a scale will be meaning that stages approved by the ISC turn into regiostages.

Ordovician (S. Finney). As the British scheme is thought to be of regional importance only, the subcommission objective is to establish the GSSP levels and sites to give corresponding names to new stages (three series and six stages are envisaged).

Silurian (Rong Jia-Yu). The task was to reconsider some GSSP, although the Silurian is the most auspicious system in terms of the global boundary stratotypes. Substantiation for eight stages of four series, which are accepted in Russian schemes as stages, is planed.

Devonian (P. Bultynck) is an auspicious system with GSSPs approved for all the stages. Seven traditional stages of three series are left as they are. It is expected that large stages (more than 10 Ma) could be divided into substages.

Carboniferous (R. Lane, representative of the subcommission); the system is divided in two the Mississippian and Pennsylvanian series. Seven stages traditional of Russian schemes (Tournaisian, Visean, Serpukhovian, Bashkirian, Moscovian, Kasimovian, Gzhelian) are likely to be left, though other versions are discussed as well. The task is to continue substantiation of GSSPs, four of which are not established.

Permian (B. Wardlaw). Subdivision into three series is proposed (four Russian stages in the lower series, three American stages in the middle, and two Chinese stages in the upper series). The Permian scale of Russia is losing three upper stages of the system.

Triassic (Chairman M. Orchard was absent); the system may be devoid of principal changes, although work on GSSPs is necessary. Two lower stages are likely to retain Russian names.

Jurassic (N. Morton); three series and eleven traditional stages are expected to be in this system. The task is to establish the desirable GSSPs (only three are authorized).

Cretaceous (P. Rawson); two series of the system include twelve traditional stages, although only three GSSPs are substantiated so far.

Paleogene (H. Luterbacher); the system is left without changes (three series and nine stages), but just a few GSSPs are admitted.

Neogene (D. Rio, representative of subcommission); except for the new Gelasian stage included into the upper Pliocene, the system underwent no other changes, although its lower part is short of GSSPs. The idea to incorporate the Quaternary into the Neogene is still under discussion. There are many pros and cons of the idea. Russian specialists are in favor of the individualized Quaternary System.

Quaternary (F. Gibbard, representative of subcommission); the appropriate Quaternary scale is not elaborated so far. The Russian scheme seems to be regarded as regional.

A.I. Zhamoida (paper in this issue) describes in detail the problems associated with the new ISS versions and outlines position of the Russian ISC with respect to many debatable points. Referring a reader to his pithy paper, I have to recall some shady moments in international activity of the ISC and National Committee of Geologists of Russia in recent years (Gladenkov, 2001).

I foresee a necessity of following organizational and scientific actions:

– We should learn to do everything in time (to discuss burning problems in the ISC, to be ready with corresponding publications, for instance on substantiation of stages etc., by the dead line, to address timely the international organizations with proposals, e.g., about the ISS structure, and the like;

– We should learn to advocate our viewpoints (and not to be guided by opinion of somebody else, or by false "patriotism") in international organizations;

– We should be certain to discuss alternative approaches in stratigraphy at the ISC meetings and to take into account the accepted recommendations in international discussions;

– We should be active and adequate participants of international stratigraphic bodies (commissions, committees, and others). We should send our representatives there in time and prepared with well-grounded opinions and proposals;

– We should learn to make decisions without haste, being guided by a scientific approach and common sense but not by modern, sometimes premature views (as it was, for instance, with the ISC decision to eliminate the Volgian Stage in the general scale, and the like);

– It should be a practice of the Russian ISC to discuss every year all urgent problems of stratigraphy (on systems, stages, regional units, etc.).

For instance, ISC could discuss now how purposeful are the "parallel" stratigraphic scales if included into the ISS (e.g., the Tethyan and Boreal schemes or marine and continental schemes). We have to get a corresponding viewpoint and particular proposals to address with to the ISC.

At the meeting in Urbino, the state-of-the-art of stratigraphic classification and nomenclature and, in particular, the comparative analysis of *International* and *Russian* stratigraphic codes were under consideration (presentation by Yu.B. Gladenkov). Omitting details, I would like to refer a reader to a recent publication (Gladenkov, 2003) elucidating the problems and basic ways of their solutions. Many Russian startigraphers acknowledge major merits of the international guide, although, they consider the idea of "multiple" stratigraphy as debatable and the missed categories of regional and other as disadvantages of the guide.

Now I dwell on the subject that is under active discussion after the ICS meeting in Urbino. This is the future of stratigraphy, as it is understood in the report of Gradstein who actually formulated a particular concept of stratigraphy development.

Gradstein pointed out that the ICS, as an efficient and authoritative organization, may exert great effect on stratigraphic studies in the world. The ICS involves the best experts who contributed much to standartization of stratigraphic units and the general scale, it is an active organizer of stratigraphic meetings and unites the most active scientists from different countries, it initiated preparation of the *International Stratigraphic Guide* and affects methods of geological mapping, etc. This does not mean however that the ICS activity was free of weak points (for instance, its certain isolation and poor connections with national stratigraphic institutions, and others). The lack of a clear stratigraphic ideology seems to be one of the main flaws (the concept of "golden nails" popular in the last years is not of course the desirable ideological basis).

In the context of new geological realities, stratigraphy faces problems somewhat shaded before. In the past half-century, problems of stratigraphic subdivision and elaboration of stratigraphic schemes and different scales were of paramount importance in stratigraphy. At present, the other aspects of stratigraphy are advanced to get a deeper insight into the essence of geological processes and their correlation (biotic evolution, climatic changes, surface and in-depth processes controlling the Earth's evolution, catastrophic events, etc.).

Consequently, it would be inadequate objective of present-day stratigraphy to complete the ISS standartization and to elaborate its modern version by 2008 (I mean to verify boundaries between all the systems, series, and stages of the global scale on the basis of recent data that will always be in progress). It is necessary to develop stratigraphy oriented to the geological process that should be its new mandate.

I omit here the particular organizational measures, which are necessary in Gradstein's opinion (e.g., structural changes in the ICS, setting up of a specialized

journal and information service, specialized stratigraphic projects, creation of database, etc.). It is necessary to point out again however that new orienting points in the ICS activity are designated and Russian stratigraphers should know them. Most likely, our specialists have already some interesting ideas, which should be properly discussed.

In Russia, new proposals received a special understanding in connection with the Vernadsky's idea of the biosphere and "former biospheres", which is popular amidst Russian geologists (Vernadsky, 1926). For Russian stratigraphers, his idea of everlasting and irreversible process in biosphere during the Earth evolution is a basis for a broad-scale ecosystem analysis of the stratisphere, its formation stages and critical events in biota history. It is significant that objects involved into analysis are of different scale, from elementary basins (basin stratigraphy) to former biospheres in general. In other words, this scientific orientation is of theoretical and practical value, as it includes stratigraphic research of individual regions and their parts. Ecostratigraphy of that kind gradually becomes capable to create the logical and methodical stability cementing stratigraphy as a whole.

Although Vernadsky's ideas are less known abroad than in Russia (his book *Biosphere* is already translated in France, USA, and other countries), many stratigraphers, following the history of science, come to similar thoughts. They feel a necessity of intent studies of historically alternating processes and phenomena reflected in stratigraphic schemes and scales. In accord with the idea, stratigraphers not only trace the diverse events recorded in particular sections, but also seek for explanation and regularities of their manifestation. In turn, this approach opens an opportunity of a deep insight into internal and cosmic organization of biosphere and into peculiar trends of biota evolution.

In this connection, a special attention will be given more and more to elaboration of real calendars of geological events (regional and global), on the one hand, and to paleogeographic and paleobiogeographic reconstructions, on the other hand, which can be regarded as visual results of "space stratigraphy."

The term "dynamic stratigraphy" suggested recently (Hallman, 1981) appears to be appropriate if we mean by that investigation of geological and biological processes and events of the past. In this understanding, the dynamic stratigraphy is able to strengthen its position in the fundamental Earth sciences, which has been lost for the past decades, as some geologists believe.

To emphasize the essence and trend of the current stratigraphic goals, stratigraphy of the 21st century should be termed "stratigraphy of biosphere" (or stratigraphy of former biospheres). In principle, dynamic stratigraphy can be regarded as part of this science.

Stratigraphy of biosphere is aimed to study fossil biospheres (former biospheres according to Vernadsky) or the state of biospheres in the past. Being a complicated and self-organizing system, biosphere consists of interconnected individual ecosystems, and the regional stratigraphy can be regarded as exploration of paleoecosystems.

The biosphere of the Earth evolved about as long as four billion years. Its evolution is driven by flows of energy, substance, and information, which are directed and regulated by the inner organization system of the biosphere. The most fundamental aspect of evolution is the evolution of the biosphere as the whole, its global organization. There are different approaches to searching for peculiar trends of the biosphere evolution and factors controlling it. Up to present, we have no general scheme of interrelation between the external and internal factors, and their impact on evolution is inadequately understood. It is evident nevertheless that stages and phases of the biosphere evolution reflect directly or indirectly the successive steps in evolution of the living substance at different levels of the Earth's biosphere and possible cosmic impacts on biota. In fact, biosphere is the most sensitive recorder of natural processes and phenomena, which occurred in the Earth's history. Therefore, "former biospheres" taken consecutively (changing of biosphere states, manifestation of biosphere rhythms) do create the basis, on which stratigraphic scales are constructed.

As noted above, stratigraphy deals with fundamental problems of general geology. First among the problems is understanding of *natural periodicity in the geological history of the Earth* (the problem is unsolvable without stratigraphy in principle). The second one it is *geological mapping*, which, no matter how changeable are methods, cannot be accomplished without stratigraphy. It is likely that both problems can be referred to everlasting. That is why stratigraphy will always be a desired science. It will become merely a more methodically furnished, more detailed, more informative in geological aspects. Thus, stratigraphy of the future needs in close connections with geologic and biologic knowledge, in more detailed, correct, and reliable methods and technology of stratigraphic analysis. Of course, we should better understand the ideological problems of stratigraphy, without which there is no sense to be involved in this science.

It should not seem strange that in addition to traditional problems of stratigraphy (stratigraphic classification, stratigraphic scales, etc.), we have to focus much attention on the *history of biosphere* and ecosystems of the past, and on different geological events. That is why the present-day stratigraphy tends to solve, besides problems of the first methodical–technological level (stratigraphic classification, establishing global or regional stratigraphic units and their boundaries, etc.), the problem of the second ideological level through the analysis of former biospheres. It turns to *interpretation of the natural periodicity in geological history of the Earth*, giving more attention to trends and successions of different-scale geological events.

As it follows from the aforesaid, stratigraphy is important for human society not only as applied science. It is of a universal methodological value. That is why it is difficult to overestimate the role of stratigraphy in bringing up the society, the young generation first of all, and its educational potential.

Over many years of the XX century, stratigraphy was regarded as the most aristocratic division of geology. I would like to believe that it would be of the same importance in the future, especially if it would turn to simultaneous interpretation of the geological past and future environmental situations on the Earth. Stratigraphy should retain its former and present-day significance for prospecting the mineral resources important for mankind (oil, gas, coal, phosphorites, and others). To sum up, I should say that a fruitful future is in store for stratigraphy of biosphere and a voluminous and interesting work for stratigraphers. This division of historical geology *will always be in demand by fundamental science.*

Reviewers M.A. Semikhatov and V.A. Zakharov

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New Permian Fusulinids from Turkey and Some Problems of Their Biogeography

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Abstract—New fusulinids from the Kinik and Beitepe localities of the Karakaya zone in Turkey are described. Representatives of genera *Pseudofusulinoides, Ultradaixina*, and *Pseudofusulina* from the first locality are presumably of the Sakmarian Age. The fusulinid assemblage of the Beitepe locality includes 23 species of 18 genera and subgenera: *Schubertella, Mesoschubertella, Minojapanella, Neimonggolina, Biwaella, Toriyamaia, Pamirina (Levenella), P. (Pamirina), Quasifusulina, Rugosofusulina, Darvasella, Dutkevitchia, Leeina, Darvasites, Robustoschwagerina, Chalaroschwagerina, Pseudofusulina*, and *Praeskinnerella.* The assemblage is typical of the Yakhtashian Stage of the Lower Permian. Biogeographic analysis showed that fusulinids of both localities mostly populated the Paleotethys northern areas, in particular, northern Afghanistan and Darvaz. Accordingly, paleobiogeographic zoning of Altiner *et al.* (2000) who included these areas in the Northern biofacies belt of the Gondwana southern margin appears deserving a criticism. New species *Pseudofusulinoides kinikensis* and *Pseudofusulina ishiklariensis* are described.

Key words: fusulinids, Permian, stratigraphy, biogeography, Turkey, Tethys.

INTRODUCTION

In Turkey, there are two broad distribution zones of Permian fusulinids, the southern one extending over the Taurus and southeastern areas of the country (Hazro and Hakkari districts), and the northern broader zone between the Aegean Sea in the west and the town of Erzincan in the east (figure). In the former zone, fusulinids occur in carbonate sequences transgressively overlying the Devonian–Lower Carboniferous sedimentary cover of the Arabian platform (southeastern Turkey) or constituting the allochthonous thrust sheets of the Taurus (Köylüoğlu and Altiner, 1989; Lys, 1986; Monod 1977; Altiner, 1984; Altiner and Özgül, 2001). In the northern zone, fusulinids are characteristic of limestones in isolated tectonic blocks set in the Triassic terrigenous–volcanogenic zone of the so-called Karakaya orogen (Erk, 1942; Şengör, Yilmaz, 1981; Leven and Okay, 1996).

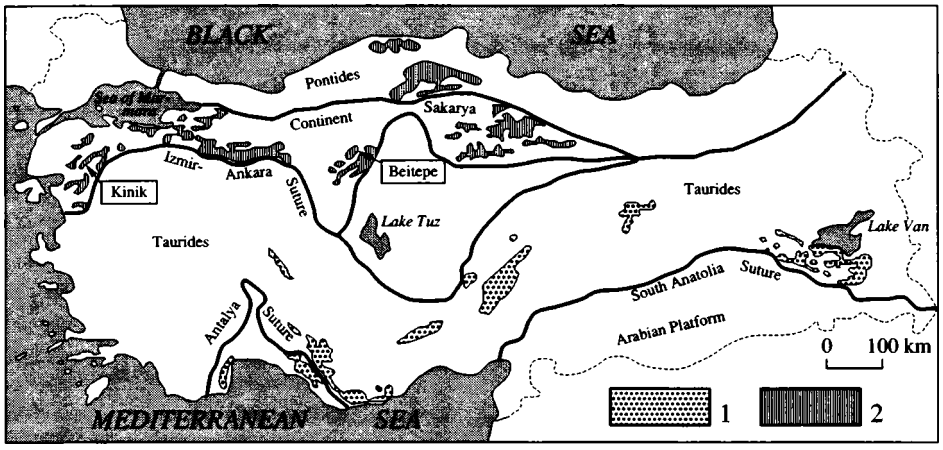
Coeval fusulinid faunas from two zones are of a contrast taxonomic diversity, consisting of taxa occurring in different proportions. Altiner *et al.* (2000) who analyzed distribution of Upper Permian fusulinids recognized a carbonate platform that rimmed the Gondwana all along its northern margin from northern Africa to Australia. Facies and faunal distinctions between the zones have been explained in terms of their internal (Southern Belt) or external (Northern Belt) disposition relative to Gondwana.

The suggested positioning of the Southern Belt relative to Gondwana seems to be reasonable, but for the Northern Belt this can be said if Turkey territory only is under consideration, for the belt eastern continuation, as it is shown in Fig. 4 in the paper by Altiner *et al.* (2000), seems to be insufficiently well substantiated in our opinion. It is quite doubtful that central Iran and entire territories of Afghanistan and Pamirs belong to the belt.

As discrimination of the Northern Belt is based on foraminiferal assemblages of the Karakaya zone, investigation and biogeographic analysis of these assemblages may be decisive by solving the problem of the belt validity. The problem has been partially discussed some time ago (Leven and Okay, 1996), and this work is a next step in that direction. We consider here two occurrence sites of Early Permian fusulinids in the Karakaya zone, one situated near the village of Beitepe close to Ankara, and the other one located southwestward of the town of Kinik (figure).

EXPOSURES AND CHARACTERIZATION OF FUSULINID ASSEMBLAGES

Beitepe locality. In 1994, one of us found Sakmarian and Yakhtashian fusulinids in limestone pebbles scattered over western slopes of the Beitepe Hills southwestward of Ankara (Leven, 1995) and assumed that pebbles are derived from rocky outcrops seen close to the hill tops. We studied one of these outcrops at the time of International symposium "PaleoForams 2001." The exposure is close to the watershed area of the



Southern (1) and Northern (2) biofacies Belts in Turkey (Altiner *et al.*, 2000).

Beitepe Hills, being seen immediately above the village build-up area. Exposed in outcrops are thick- (below) and medium-bedded (higher) algal and algal-fusulinid micritic limestones of biodetrital type. The beds' attitude is almost horizontal or slightly dipping eastward. The beds are about 20 m thick in total. Below the outcrop, the slope is covered by vegetation, and underlying rocks remain unknown.

We collected 16 samples from the limestone section and found fusulinids in many of them. A clear trend in distribution of fusulinids has not been established, and we present below only the list of all the taxa identified (species occurring in the Darvaz region are marked by asterisks).

These are *Schubertella simplex* Lange*, *Schubertella* sp. indet., *Mesoschubertella thompsoni* Sakagami*, *Mesoschubertella* sp., *Minojapanella elongata* Fujimoto et Kanuma*, *Neimonggolina* cf. *fusiformis* Xia, *Pamirina (Levenella) leveni* Kobayashi*, *P. (Pamirina) darvasica* Leven*, *Pamirina (Pamirina)* sp., *Biwaella europae* Kochansky-Devidé et Milanovich*, *Quasifusulina magnifica* Leven*, *Rugosofusulina netchaevi* Leven et Scherbovich*, *R. mariae* Leven et Scherbovich*, *Rugosofusulina* sp., *Darvasella vulgariformis* (Kalmykova)*, *D. compacta* (Leven)*, *D.* aff. *priva* Leven*, *Dutkevitchia jipuensis* (Nie et Song), *Leeina kraffti* (Schellwien et Dyhrenfurth)*, *Darvasites contractus* (Schellwien et Dyhrenfurth)*, *D. ordinates* (Chen)*, *D. simplex* (Zhou), *Darvasites* sp., *Robustoschwagerina* sp.*, *Chalaroschwagerina vulgaris* (Schellwien et Dyhrenfurth)*, *Ch.* cf. *darvasica* Leven*, *Chalaroschwagerina* sp., *Pseudofusulina* (?) *hunanensis* (Zhou), and *Praeskinnerella cushmani* (Chen).

About 500 m to the north, there is a roadside excavation, where marly shale and marl with thin interlayers of micritic fusulinid limestone are exposed. Relationships between these rocks and limestones described above are unclear, although the former appear to be exposed at a lover hypsometric level. *Pseudofusulinoides* forms found here belong to the new species *Pseudofusulinoides kinikensis* from the Kinik locality characterized below.

Kinik locality. This locality near the village of Işiklar is situated 10 km southwestward off the town of Kinik (figure). One of us observed outcrops here during excursion organized by Prof. A. Okay from the Technological University of Istanbul. In opinion of Okay, in the upper part of a thick sequence of Triassic sandstones and shales exposed in this area there are large olistoliths of limestone attributed to the Hodul Beds (Leven and Okay, 1996), two of which have been studied.

In one olistolith, there are seen beds of dark-colored layered quartz siltstone and sandstone about 150 m thick in total, having the unclear poorly exposed contact with overlying bed of yellowish foliated marl (1 m) that grades upward into light gray biomicritic limestone 2 to 3 m thick. The higher interval of slope about 10 m wide is covered with vegetation, and dark limestone and sandstone beds of the next interval bear *Eopolydiexodina.* The sequence cannot be traced farther with confidence, because the rocks are poorly exposed and crushed. At the base of marl, there is an interlayer (3–5 cm) of gastropod coquina overlying, as one can judge from rock detritus, the siltstone and sandstone beds. The contact proper is unexposed, though it seems to be of a stratigraphic type. The marl bed yields fusulinids, which are represented by the new species *Pseudofusulinoides kinikensis* only. Among abundant fusulinids of the overlying limestone, we identified *Ultradaixina postgalloway* (Bensh) and *Pseudofusulina ishiklariensis* n. sp.

The second olistolith is exposed not far away near the village of Çamoba. In its lower part exposed above the slope interval covered with vegetation, fusulinid limestones are intercalated with beds of sandstone with interlayers of gravelstone and small-pebbled conglomerate. The apparent thickness of rocks is not greater than few dozens of meters. Predominant among fusulinids are the late Murgabian–Midian *Eopolydiexodina, Neoschwagerina,* and *Afghanella* forms. It is likely that limestone beds, which are exposed in the

previous outcrop above limestone with *Ultradaixina* and *Pseudofusulina*, belong to this stratigraphic interval. These rocks are separated by a distinct, though stratigraphic contact from the overlying sequence (about 100 m thick) of light-colored compact to flaggy micritic limestones of deep-water origin, which yield the Dzhulfian–Dorashamian *Palaeofusulina (Paradunbarula), Reichelina*, and smaller foraminifers. The Murgabian–Midian and Dzhulfian–Dorashamian fusulinids from both outcrops and other olistoliths of the Karakaya zone have been described already (Erk, 1942; Leven and Okay, 1996).

Published data and new results of observations in the Beitepe and Kinik localities allow us to suggest the following general succession of Permian beds in the carbonate platform, fragments of which are incorporated into the Karakaya complex (from the base upward):

(1) Siltstones and sandstones of the Kinik locality whose age is unclear; Okay attributed these rocks to the Norian Stage based on *Halobia* specimens found in similar sandstones and shales of the adjacent area (Leven and Okay, 1996). The presumable stratigraphic contact between them and overlying fusulinid limestones implies that the rocks are of the Permian age at any rate, not older. We do not exclude, however, that a comparable terrigenous rocks may be present in the Triassic sequence as well.

(2) Marls quickly grading upward into micritic limestones with *Pseudofusulinoides* and *Ultradaixina*; their contact with underlying siltstones is likely stratigraphic, although a hiatus in sedimentation is admissible for this level. The limestones belong to the Sakmarian Stage (?).

(3) Biogenic limestones of the Beitepe locality, where they yield Sakmarian and Yakhtashian fusulinids (*Zellia, Robustoschwagerina, Pamirina, Chalaroschwagerina, Darvasites, Praeskinnerella*); relationships between these and underlying rocks are unclear (Leven, 1995; and data of this work).

(4) A problematic Bolorian–Kubergandian interval of the succession; the Kubergandian fusulinids (*Parafusulina, Cancellina, Armenina*) are already known from several outcrops of the Karakaya zone (Leven and Okay, 1996), but Bolorian taxa have not been found so far.

(5) The Murgabian–Midian fusulinid limestones; the rocks yield *Eopolydiexodina, Neoschwagerina, Yabeina, Afghanella, Sumatrina*, and *Verbeekina* forms (Erk, 1942; Skinner, 1969; Leven and Okay, 1996).

(6) Light-colored compact micritic limestones of the Kinik locality, which yield *Palaeofusulina (Palaeofusulina), P. (Paradunbarula)*, and *Reichelina* species of the Dzhulfian–Dorashamian age; the rocks conformably rest on underlying strata (Leven and Okay, 1996).

It is remarkable that a greater portion of the rock sequence is composed of shallow-water facies in distinction from limestones of Unit 6, which were deposited in a deeper setting. Kozur and Kaya (1994) who studied conodonts already paid attention to the basin deepening since the Dzhulfian Age. This conclusion is consistent with known data on Transcaucasia and Iran (Abadeh site), where the Midian–Dzhulfian boundary also marks a transition from the shallow- to deeper-water facies. The pelagic fauna substitutes the benthic one across the same boundary (Iranian-Japanese..., 1981; Leven, 1998).

ANALYSIS OF FUSULINID ASSEMBLAGES

Fusulinid assemblages of the Karakaya zone had been already characterized in general (Leven and Okay, 1996). Accordingly, it seems reasonable to consider here only fusulinids from the Kinik and Beitepe localities.

The fusulinid assemblage of the *Kinik locality* includes three genera only, two of which are represented therewith by new species unsuitable for determination of the host rock age. The genus *Pseudofusulinoides* has been described for the first time from Asselian deposits of the southern Fergana (Bensh, 1972). It seems to be more characteristic of the Sakmarian Stage however, since representatives of the genus are found in deposits of that stage in the Carnic Alps (Forke, 1995), northern Afghanistan (Leven, 1971), and Xinjiang (Chang, 1963). We should note in addition that *Pseudofusulinoides* forms from the Kinik locality are, as it follows from the large size of shells having many whorls, greatly advanced in the evolutionary aspect.

Species *Ultradaixina postgalloway* (Bensh) was found for the first time in upper beds of the Uluk Formation of northern Fergana, which discordantly overlies the Asselian deposits. Based on rocks' position in the section and on discovered fusulinids, Bensh (1962) correlated the Uluk Formation with Sakmarian strata of southern Fergana and with deposits of the Yakhtashian Stage in the Darvaz region. V.I. Davydov substantiated a considerably greater stratigraphic range of that species, because it was established in boundary strata separating the Upper Carboniferous and Asselian deposits in the Darvaz region and southern Urals (Chuvashov *et al.*, 1986). In the Carnic Alps, the species was found at the same stratigraphic level (Forke *et al.*, 1998). Thus, the age of deposits bearing *U. postgalloway* can be defined with a low precision, only within a broad interval from the Late Carboniferous to the Yakhtashian deposits of the Early Permian. However, species from the Kinik locality and type specimens from the northern Fergana are larger than the Late Carboniferous counterparts and have a more distinct and regular plication of septa. In accord with opinion of Bensh, they seem to be younger, likely Sakmarian in age.

The genus *Pseudofusulina* found in the Kinik locality is widespread, having a wide stratigraphic range. In our collection, it is represented by new species *Pseudofusulina ishiklariensis* resembling *Pseudofusulina*

forms from the upper Wolfcampian–lower Leonardian of America (Thompson, 1954; Dunbar and Skinner, 1937) and *P. ambiqua* (Deprat) occurring in Yakhtashian–Bolorian deposits of Indochina, Japan, China, and Darvaz (Deprat, 1913; Leven *et al.*, 1992).

The more diverse fusulinid assemblage from the *Beitepe locality* is typical of the Yakhtashian Stage. Among 23 species listed higher by the locality description, 18 taxa are characteristic of the Darvaz type sections of the stage. Many of them occur in all other Tethyan sections. Only four of 23 species have not been found in the Darvaz region so far. These are *Neimonggolina fusiformis, Dutkevitchia jipuensis, Darvasites simplex*, and *Praeskinnerella cushmani*. The first one is known however from Inner Mongolia, where it occurs in association with Kubergandian fusulinids (Xia, 1982). The second species is a component of the Bolorian(?)–Kubergandian fusulinid assemblages from the southern Tibet (Nie and Song, 1983) and southern Afghanistan (Leven, 1997). The other two forms were first detected in Yakhtashian deposits of southern China (Chen, 1934; Zhou, 1982), and similar forms are characteristic as well of coeval deposits in the northern Afghanistan, Darvaz, and Xinjiang.

Genera and species of the assemblage under consideration occur beyond the Darvaz also in the Saidi Kajon section of northern Afghanistan (Leven, 1997). Somewhat modified their assemblages are known as well from several sections of southern China and Japan.

SOME BIOGEOGRAPHIC CONCLUSIONS

Our earlier analysis of fusulinid assemblages from the Karakaya zone (Leven and Okay, 1996) showed that, being different from assemblages of southern and southeastern Turkey (Southern Belt after Altiner), they are nevertheless very similar to assemblages of northern Afghanistan, northern Pamirs, and Darvaz, i.e., of areas characterizing the northern peri-Eurasian margins of the Paleotethys. New data considered above are consistent with this conclusion and contradict the idea that the Northern Belt, where the Karakaya zone is located, extended all around the Gondwana outer margin (Altiner *et al.*, 2000). Presented below is our attempt to clarify the idea of two peri-Gondwanan biofacies belts based on all the data concerning the problem.

Discriminating the belts, Altiner and his colleagues analyzed the Late Permian biofacies only. They included the Taurus, the Arabian platform margin, southern Iran, Saudi Arabia, Oman, West Yunnan province, Burma (Myanmar), and Thailand in the Southern Belt. The Northern Belt incorporates, according to their idea, the Karakaya zone of Turkey, Crimea, Transcaucasia, central Iran and Elburz, Afghanistan, Pamirs, Tibet, northern and southern China, Indochina, and Japan (Altiner *et al.*, 2000). The main discrimination criteria used in the cited work were the distinctions in abundance and diversity of Late Permian fusulinid assemblages of two belts, as fusulinids of the Northern Belt were found to be more abundant and diverse than in the Southern Belt. In addition, the authors of idea took into consideration the distribution areas of smaller foraminifers representing genera *Paradagmarita* and *Shanita*, which occur in the Southern Belt only.

With these criteria taken into consideration, we cannot understand why central Iran, Elburz, and Transcaucasia should be included into the Northern Belt. In abundance and taxonomic diversity of fusulinids, the Upper Permian sections of these regions are comparable rather with sections of the Taurus than with those of the Karakaya zone. In particular, this is correct with respect to sections of central Iran and Elburz, where the Jamal and Rute formations of the Upper Permian are practically barren of fusulinids (Jenny-Deshusses, 1983; Leven and Vaziri, in press). These fossils are somewhat more abundant in the Abadeh section of central Iran and in Transcaucasia (Iranian-Japanese..., 1981; Leven, 1998), although even there fusulinids occur occasionally in thin interlayers intercalated with thick limestone beds, which are barren of this fauna. The same distribution trend is typical of the majority of sections in southern Afghanistan, southern Pamirs (Leven, 1967, 1993, 1997), Karakorum, Salt Range, Himalayas, and southern Tibet (Gaetani *et al.*, 1995; Pakistan-Japanese..., 1985; Sun, 1993). Accordingly, the indicated sections belong rather to the Southern than to the Northern Belt, and this is consistent with geographic distribution of smaller foraminifers as well. For instance, the genus *Shanita* characteristic of the Southern Belt is known from sections of southern Afghanistan (Leven, 1997), southern Pamirs (Leven, 1991) and from the Quiangtang Block of southern Tibet (Sun, 1993). Similarly characteristic of the peri-Gondwanan areas of the Tethys only is the genus *Sphairionia* of smaller foraminifers, which occurs, beyond the Taurus, in Transcaucasia, central Pamirs, Kampuchea, Malaysia, and western Thailand (Pronina, 1995). Tight connections between Gondwana and these regions are clearly evident from data on distribution of brachiopods, bivalves, bryozoans, and other faunas (Grunt and Dmitriev, 1973; Goryunova, 1975; Termier *et al.*, 1973, 1974; Angiolini, 1994). As for the Northern Belt, its fusulinid assemblages from the Karakaya zone are very similar, as mentioned earlier, to assemblages of northern Afghanistan and Darvaz, and, according to present-day reconstructions, both regions were in the Permian time closer to the Eurasian coast of Paleotethys, than to the coast of Gondwana.

The same conclusion is inferable from biogeographic data on the Early Permian fusulinids. The Sakmarian–Yakhtashian fusulinids of the Karakaya zone studied in this work are unambiguously correlative with coeval assemblages of northern Afghanistan, Darvaz, and Fergana, i.e., of the regions that neighbored Laurasia. Similar fusulinid assemblages are known in the Quen Lun, Cinling, South China, Indochina, and Japan (Zhao *et al.*, 1984; Wang and Sun, 1973; Xiao *et al.*,

1986; Igo *et al.*, 1993; Morikawa and Isomi, 1961). In the peri-Gondwanan areas, they are practically unknown, because some Yakhtashian forms have been reported from southern Tibet only (Nie and Song, 1983). Thus, like in the Late Permian, the peri-Gondwanan areas of the Early Permian time were distinguishable from the other Tethyan areas. Even if the Southern biofacies belt of peri-Gondwanan areas is somehow admissible for the Late Permian, because corresponding carbonate facies are depleted in fusulinids everywhere in these areas, nothing of this kind can be said about the Early Permian. One can actually observe the Lower Permian (primarily Asselian–Sakmarian) carbonate facies in southern Turkey and central Iran, but further to the southeast, along the Gondwana margin (southern Afghanistan, southern Pamirs, Karakorum, southern Tibet, western Indochina), they become replaced by terrigenous deposits. The character of Early Permian biota changes in accord with this trend: the Asselian fusulinids are known only in sections of the Taurus and central Iran, while they have not been found farther eastward. The Sakmarian fusulinids are present in the Taurus and, presumably, in central Iran (Abadeh site). They occur as well in sections of southern Afghanistan, central Pamirs, Karakorum, and southern Tibet, but are represented in these regions by the endemic Kalaktash fauna of a low diversity (Leven, 1993, 1997; Gaetani *et al.*, 1995; Nie and Song, 1983). Further to the east, Sakmarian fusulinids disappear. As was stated already, assemblages of typical Sakmarian fusulinids have not been encountered in peri-Gondwanan areas. The Bolorian fusulinids occurring in central Iran, Transcaucasia, southern Afghanistan, southern Pamirs, and Karakorum are unknown from areas located further eastward. Thus, the outlined geographic distribution of sedimentary facies and their fusulinid assemblages discredit the suggested existence of the Southern biofacies belt that united peri-Gondwanan areas during the Early Permian.

Accordingly, two biofacies belts discriminated by Altiner and his colleagues appear to be inconsistent with data on distribution of foraminifers. Taking fusulinid assemblages of the Karakaya zone for a premise, one should conclude that the "Northern Belt" must include the areas, which were located in the northern periphery of the Paleotethys (Carnic Alps, northern Afghanistan, northern Pamirs, Fergana region, Xinjiang, Quen Lun, Chinling, northern to southern China, Japan, and the greater area of Indochina), but not the regions of central Iran, southern Afghanistan, southern Pamirs or others listed above. Both the Early and Late Permian fusulinid assemblages from the northern periphery of the Paleotethys are well correlative with their counterparts from the Karakaya zone. In such a case, the "Northern Belt" would be covering almost the entire area of the Paleotethys, and idea of the "belt" is loosing a sense therefore. In general, considering all the data on the Permian Tethys, we think that it would be more correct to speak of North and South Tethyan biogeographic provinces, but not of two peri- Gondwanan "biofacies belts," and that distinctions between the provinces were controlled by climatic factors (Leven and Shcherbovich, 1978) and, probably, by incipient individualization of the Neotethys.

Summarizing data on biogeography of fusulinids, we can state that they populated the Paleotethys everywhere from the Mediterranean areas to the Far East. Characteristics of their communities unchangeable in general varied therewith depending on local facies environments. A good example is the order Neoschwagerinida similarly represented everywhere except for the subfamily Lepidolininae confined to the Far East areas of the Paleotethys, while the family Sumatrinidae was mainly widespread in western areas (Kobayashi, 1997, 1999). The fusulinid community in a basin that existed in the Karakaya zone was typical of the Paleotethys and consequently communicated with main biota of this Permian ocean.

Epicontinental basins of southern Turkey were situated in the northern periphery of Gondwana, being more or less isolated from the main water mass of the Paleotethys in contrast to the Karakaya basin that easily communicated with the latter. Isolation and depressed water circulation could produce an impact on salinity parameters in southern basins, creating unfavorable habitat environments for stenohaline organisms like corals and fusulinids. The latter penetrated in these basins during the short-term transgression episodes, when environments became normal. Such a scenario is appropriate for basins of central Iran and Transcaucasia. Paleobiogeography of peri-Gondwanan basins located further eastward was likely under a considerable influence of climatic factors, because the coastal line of Gondwana trended southeastward crossing paleolatitudes, and remoteness of basins from equator was increasing therewith. The impact of climatic factor was most perceptible in the first half of the Early Permian, when climate was contrasting and terrigenous sedimentation inhibited deposition of carbonates. As a result, thermophilic fusulinids are either missing from sediments of that period (Asselian Age), or represented by endemic assemblages (Sakmarian Age). This is observable in sections of southern Afghanistan, southern Pamirs, Karakorum, Salt Range, Kashmir, southern Tibet, Yunnan (Baoshan block), and western Thailand (Leven, 1967, 1993, 1997; Gaetani *et al.*, 1991, 1995; Pakistan-Japanese..., 1985; Nakazawa *et al.*, 1975; Sun, 1993; Wang *et al.*, 2001). In the Late Permian, especially during the Midian Age, environments were more or less close to normal because of global warming and vast transgression. Accumulation of carbonate deposits replaced the terrigenous sedimentation, and fusulinids penetrated into Gondwanan shelf seas from northern warm areas of the Paleotethys. The Late Permian fusulinid assemblages of the peri-Gondwanan basins did not differ in principle from fusulinid faunas of the Paleotethys, although they were less diverse, especially in southernmost (Salt Range, Himalayas, southern

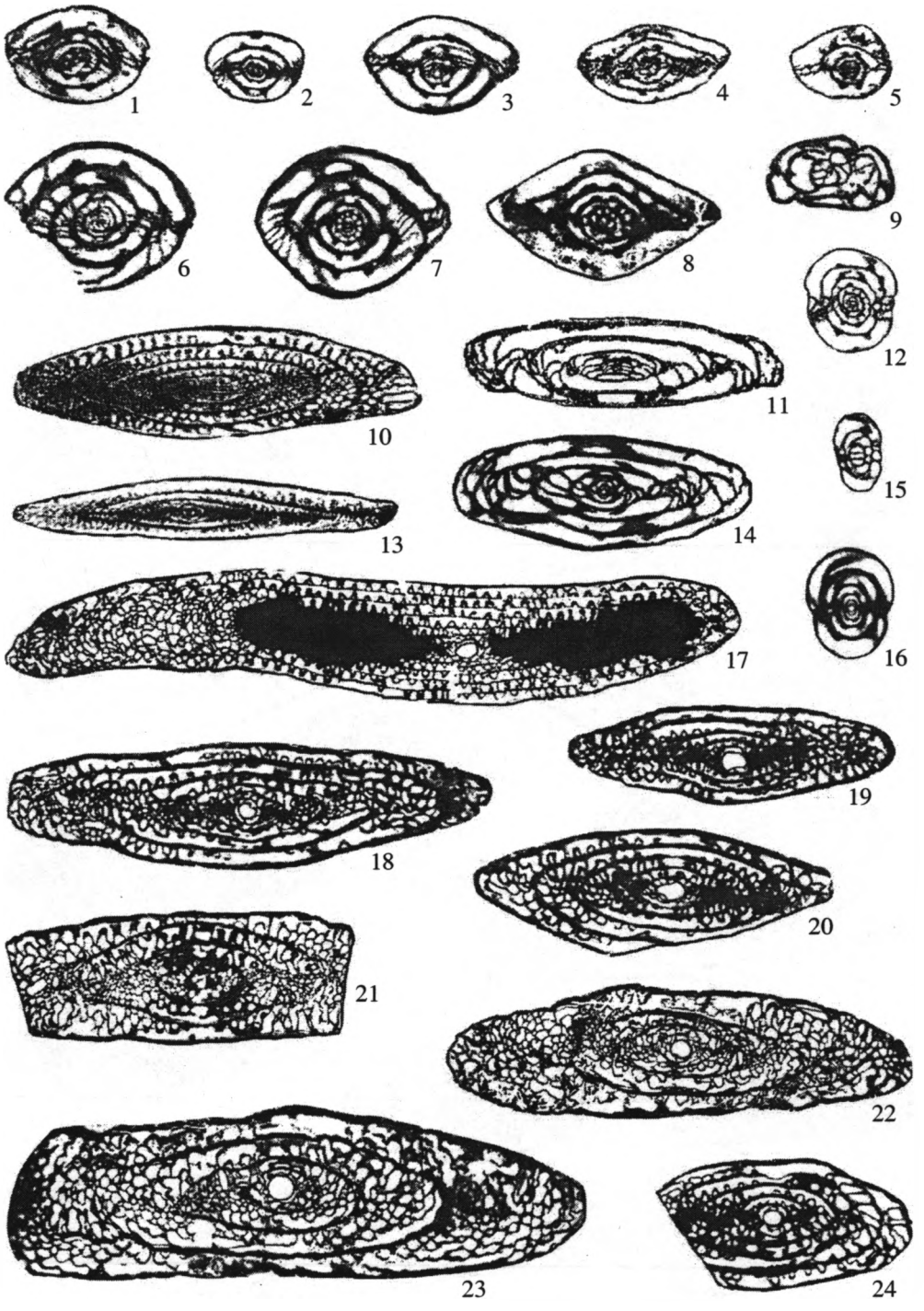


Plate I. Yakhtashian fusulinids from the Beitepe locality.

(1–3) *Schubertella simplex* Lange, ×22: (1, 3) subaxial sections, GIN 4778/1 and GIN 4778/3, (2) axial section, GIN 4778/2; (4, 5) *Schubertella* sp. ×22, subaxial sections, GIN 4778/4 and GIN 4778/5; (6, 7) *Mesoschubertella thompsoni* Sakagami, ×22, subaxial sections, GIN 4778/6 and GIN 4778/7; (8) *Mesoschubertella* sp. ×22, axial section, GIN 4778/8; (9) *Toriyamaia* sp. ×15, tangential section; (10) *Neimonggolina* cf. *fusiformis* Xia, ×22, subaxial section, GIN 4778/10; (11) *Biwaella* sp. ×15, subaxial section, GIN 4778/11; (12) *Pamirina (Pamirina)* sp. ×22, axial section, GIN 4778/12; (13) *Minojapanella elongata* Fujimoto et Kanuma. ×22, axial section, GIN 4778/13; (14) *Biwaella* aff. *europae* Kochansky-Devide et Milanovich, ×15, subaxial section, GIN 4778/14; (15) *Pamirina (Levenella) leveni* Kobayashi, ×22, subaxial section, GIN 4778/15; (16) *Pamirina (Pamirina) darvasica* Leven, ×22, axial section, GIN 4778/16; (17) *Quasifusulina magnifica* Leven, ×7.3, axial section, GIN 4778/17; (18, 19) *Rugosofusulina netchaevi* Leven et Scherbovich, ×7.3, axial sections, GIN 4778/18 and GIN 4778/19; (20, 24) *Rugosofusulina mariae* Leven et Scherbovich, ×7.3, axial sections, GIN 4778/20; (21) *Dutkevitchia jipuensis* (Nie et Song), ×7.3, subaxial section, GIN 4778/21; (22, 23) *Darvasella compacta* (Leven), ×7.3, axial sections, GIN 4778/22 and GIN 4778/23.

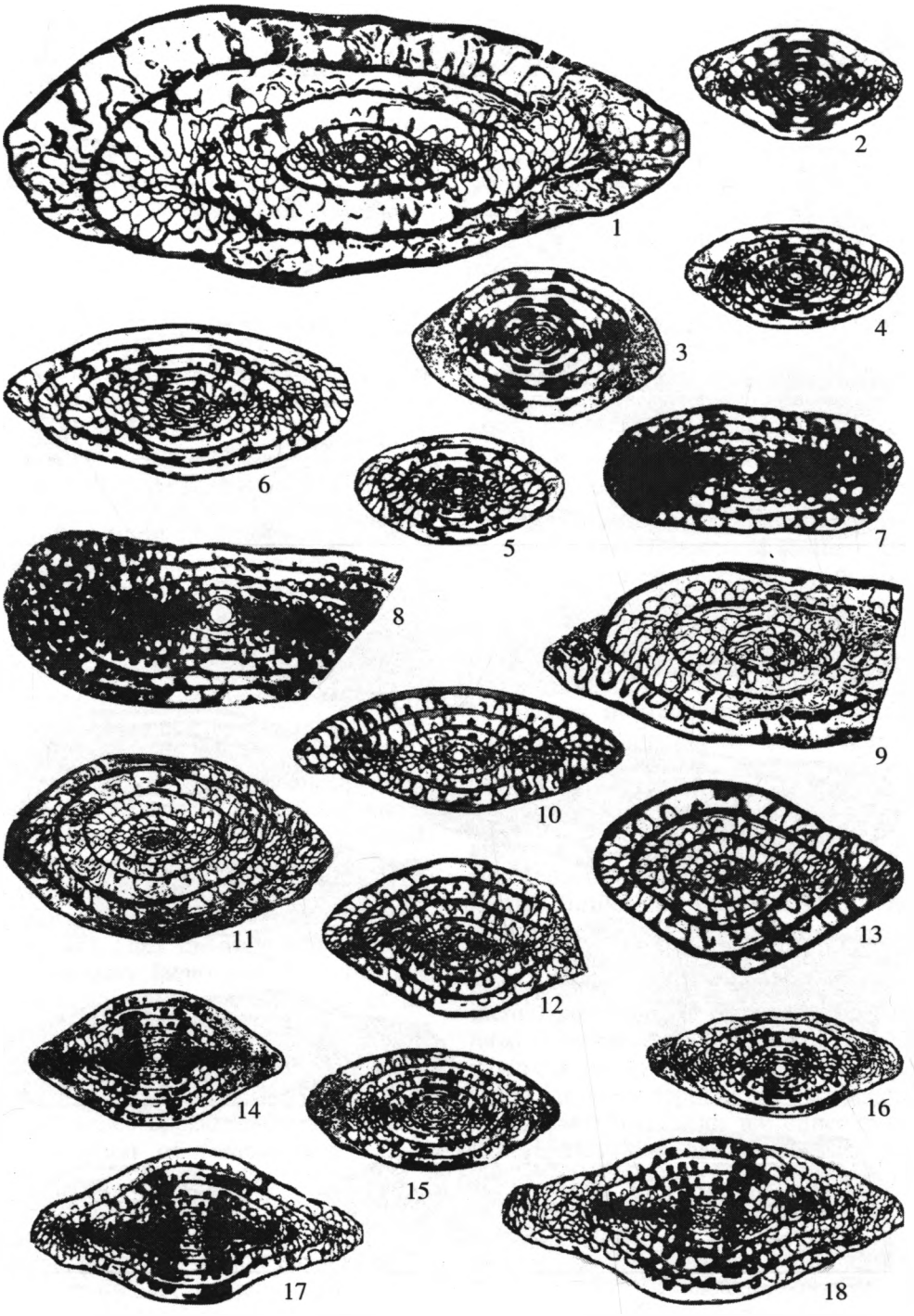


Plate II. Yakhtashian fusulinids from the Beitepe locality (magnification ×7.3 for all the figures).

(1) *Darvasella* aff. *priva* Leven, subaxial section, GIN 4778/24; (2, 3) *Darvasites ordinates* (Chen), axial sections, GIN 4778/25 and GIN 4778/26; (4, 5) *Darvasites contractus* (Schellwien et Dyhrenfurth), subaxial sections, GIN 4778/27 and GIN 4778/28; (6) *Darvasites* aff. *vozginensis* Leven, subaxial section, GIN 4778/29; (7, 8) *Leeina kraffti* (Schellwien et Dyhrenfurth), axial sections, GIN 4778/30 and GIN 4778/31; (9) *Chalaroschwagerina* cf. *darvasica* Leven, subaxial section, GIN 4778/32; (10) *Pseudofusulina hunanensis* Zhou, axial section, GIN 4778/33; (11) *Chalaroschwagerina* sp., subaxial section, GIN 4778/34; (12, 13) *Chalaroschwagerina vulgaris* (Schellwien et Dyhrenfurth): (12) axial section, GIN 4778/35, (13) subaxial section, GIN 4778/36; (14, 17, 18) *Praeskinnerella cushmani* (Chen), axial sections GIN 4778/37, GIN 4778/40 and GIN 4778/41; (15, 16) *Darvasites simplex* (Zhou), axial sections, GIN 4778/38 and GIN 4778/39.

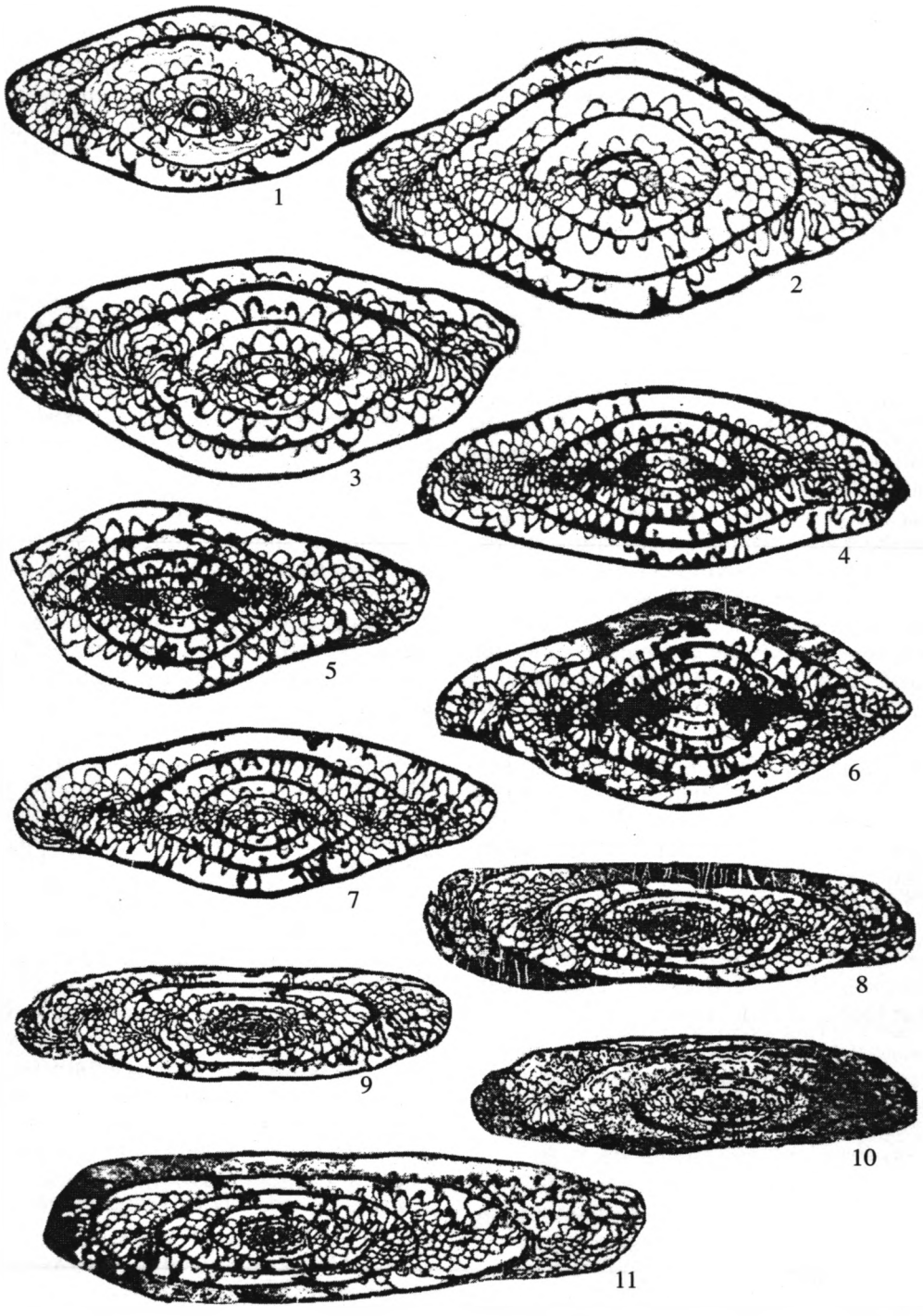


Plate III. Sakmarian (?) fusulinids from the Kinik locality, except for specimen in fig. 10 (magnification ×7.3 for all the figures). (1–3) *Ultradaixina postgalloway* (Bensh), axial sections, GIN 4778/42, GIN 4778/43 and GIN 4778/44; (4–7) *Pseudofusulina ishiklariensis* Leven, n. sp.: (4) axial section of holotype, GIN 4778/45, (5–7) axial sections of topotypes, GIN 4778/46, GIN 4778/47 and GIN 4778/48; (8–11) *Pseudofusulinoides kinikensis* Leven n. sp.: (8) axial section of holotype, GIN 4778/49, (9, 11) subaxial sections of topotypes, GIN 4778/50 and GIN 4778/52, (10) axial section, Beitepe locality, GIN 4778/51.

Tibet) and eastern regions (Baoshan block). We discussed only the most general peculiarities of biogeography of Permian fusulinids. In our opinion, biogeographic zoning was likely much more complex, depending on trends of sedimentation in various basins, which changed their configuration, and on many other factors, e.g., on seaways connecting the basins, which opened and closed in a capricious order. The under-

standing of variable paleogeography of the Permian time is an intriguing objective for stratigraphers, although all the related questions can hardly be answered in the study region of a very complex tectonic structure.

CONCLUSION

We deduce the following inferences from data considered in this work:

(1) The idea of two the Southern and Northern biofacies belts is applicable for the territory of Turkey only. The Early and Late Permian fusulinid assemblages of the Karakaya zone are correlative with coeval assemblages from all other areas of the Paleotethys, including the regions, which were situated in the northern periphery of that paleoocean. The notion "Northern Belt" is loosing a sense therefore.

(2) To speak of a continuous Southern biofacies belt is possible for the Late Permian only, mainly for the Midian Age, when carbonate sedimentation dominated everywhere in peri-Gondwanan self seas. In the Early Permian basins, sedimentation and biofacies environments changed considerably by advancing along the Gondwana margin from the northwest to the southeast. Consequently, an integral "biofacies belt" did not exist at that time.

(3) If there was the Southern Belt of the Late Permian time, then, based on indicative criteria suggested by Altiner and his colleagues, one should include in this belt the territories of central Iran, Transcaucasia, southern Afghanistan, southern Pamirs, southern Tibet, southwestern Yunnan (Baoshan block), a part of Myanmar, western Thailand, and western Malaya (Sibumasu block), in other words, all the peri-Gondwanan areas of the Tethys. Altiner *et al.* included all these areas, except for the Baoshan block, in their Northern Belt.

DESCRIPTION OF NEW FUSULINID SPECIES

Genus *Pseudofusulinoides* Bensh, 1972

Pseudofusulinoides kinikensis Leven, n. sp.

Plate Ill, figs. 8–11

Holotype: GIN 4778/49, axial section; Turkey, Kinik; Lower Permian, Sakmarian (?) Stage.

Material: 5 axial and subaxial sections.

Description. Large test fusiform in first five whorls and subcylindrical, with rounded ends in the last two–three whorls; low spiral is regularly expanding. Adult specimens have 8–9 whorls: L = 8.5–12 mm, D = 2.3–3.0 mm, L/D = 3.5–4. Wall consists of tectum and well-seen keriotheca, being 0.08–0.1 mm thick in the last whorls. Septa are plicated from pole to pole of test; septal plicae low and rounded in the middle part turn into axial fine-meshed interlacements; phrenotheca present. Low and narrow cuniculi are seen in outer whorls. Embryonic chamber small and spherical with external diameter 0.12–0.16 mm; tunnel is narrow in inner whorls and broad afterward, changing position inside whorls. Minor chomata are visible in embryonic chamber and in next two–three whorls.

Comparison. In distinction from other *Pseudofusulinoides* species, *Pseudofusulinoides kinikensis* n. sp. is larger, demonstrating quick elongation of test in two–three last whorls.

Locality and age: Kinik and Beitepe sections in Turkey; Lower Permian, Sakmarian (?) Stage.

Genus *Pseudofusulina* Dunbar et Skinner, 1931

Pseudofusulina ishiklariensis Leven, n. sp.

Plate III, figs. 4–7

Holotype: GIN 4778/45, axial section; Turkey, Kinik; Lower Permian, Sakmarian (?) Stage.

Material: 6 axial sections.

Description. Relatively large fusiform test with straight or slightly concave sides; axial ends are smoothly rounded; whorl number in adult specimens is 5 to 5.5: L = 9–10.5 mm, D = 3.5–4.2 mm, L/D=1.9–2.9. Wall consists of tectum and distinct alveolar keriotheca, being 0.12 mm thick in the fifth whorl; indistinctly seen phrenotheca present. Thin septa are regularly plicated along the test length; septal plicae are high and narrow all along septa width. Outer diameter of spherical embryonic chamber is 0.22–0.35 mm; tunnel low and narrow in inner whorls is getting wider afterward; minor chomata are present in embryonic chamber only. Small axial thickenings are visible in inner whorls.

Comparison. In distinction from *Pseudofusulina ambiqua* (Deprat), *Pseudofusulina ishiklariensis* n. sp. resembling it in elongated shape is larger, more swollen in the middle part, showing presence of phrenotheca. Characteristic of new species is as well a certain similarity with species *Pseudofusulina hessensis* (Dunbar et Skinner) and *P. neolata* (Thompson) from American localities. As compared to first of them, tests of new species are more swollen, loosely spiraling, displaying less intense and more regular septa plication. In distinction from tests of *Pseudofusulina neolata*, they are less densely coiled, having lesser number of whorls, and larger in size.

Localities and age: the same as for holotype.

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The Thermal Regime Dynamics in Callovian–Oxfordian Seas of Northwestern Eurasia: Implications of Relative Paleotemperature Data

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Abstract—Six paleothermal and paleohydrologic settings in Callovian–Oxfordian seas of northwestern Eurasia are modeled for six time intervals corresponding in duration to the substage. The modeling principle is based on the study of probabilistic relationships between various faunal elements from different paleobiogeographic provinces in a single species assemblage, ammonites in the considered case. The method opens a possibility to recognize periods of subglobal and local warming and cooling, to reconstruct currents in the entire sea system, and to specify paleotemperature distribution. Two periods of water mass warming (middle Callovian and middle Oxfordian) and three periods of their relative cooling (early Callovian, Callovian–Oxfordian boundary period, and late Oxfordian) are distinguished in the Callovian–Oxfordian. All paleobasins of northwestern Eurasia (Arctic, Central Russian, European, Peri-Tethyan, Tethyan) had autonomous circulation systems with different paleotemperature characteristics. Because of this, dynamics of the thermal regime in the Callovian–Oxfordian basins primarily reflects variations in the hydrological regime rather than paleoclimatic changes.

Key words: Callovian, Oxfordian, paleobiogeography, northwestern Eurasia, paleotemperatures, hydrological regime.

INTRODUCTION

During the Callovian–Oxfordian ages of the Middle–Late Jurassic, the territory of northwestern Eurasia was covered by vast epicontinental seas, which are traditionally subdivided into the European and Central Russian basins separated by the Scandinavian massif and Ukrainian land. In the north, the seas joined the Boreal (Arctic) basin via the Pechora seaway of the Central Russian sea and via a narrow strait between Greenland and Scandinavian massif (European sea). In the south, both basins, the boundary between which should probably be placed along the southern margin of epi-Hercynian massifs, were bordered by warm waters of the Tethys.

Although the Jurassic marine sequences in the territory of northwestern Eurasia are well studied, the paleotemperature characteristic of relevant seas remains poorly known so far. Until the appearance of paleothermometry methods, this aspect of paleobiogeographic studies was virtually ignored. The majority of authors, who were dealing with the problem of sea paleotemperatures, considered it as a problem of paleoclimatology and used the corresponding terminology and conceptual arsenal. Inasmuch as the methods of absolute paleothermometry are applicable only for determination of water paleotemperature, the relevant problems go however far beyond paleoclimatology and require the other, at least paleohydrological conceptual basis. In his fundamental work *Jurassic Geology of the World*, Arkell (1961) does not use the notion "paleotemperature" at all and related aspects are discussed in small chapter dedicated to the Jurassic climate. He emphasized therewith that even this chapter could be excluded from the book because of absence of reliable data. Discussing paleotemperatures in Jurassic seas, Arkell used relative characteristics such as indicators of warm, tropical and subtropical seas. Among the mentioned indicators, there were reef-building corals, large mollusks, and prevalence of carbonate facies in studied lithofacies. He also compared data on marine sediments with observations in continents to reconstruct the Jurassic climate. This approach allowed him to arrive at the following conclusions: (1) the Jurassic climate was warmer than now; (2) equator was substantially closer to the north, approximately in the southern part of the modern Mediterranean region; (3) characteristic of the Jurassic were alternating cooling and warming periods with a maximal warming in the Oxfordian. The last inference was derived mainly from distribution of corals and thermophilic ammonite assemblages in marine sediments. Summarizing, Arkell assumed that solar radiation in the Jurassic was higher than now.

With the paleothermometry method proposed by H.C. Urey in the 1950s, absolute paleotemperatures can be estimated based on investigation of marine organo-

genic carbonates. During subsequent decades, numerous works, summarizing included, were dedicated to Jurassic paleotemperatures of northern Eurasia and other regions (Emiliani, 1967; Bowen, 1969; Stevens, 1971; Stevens and Clayton, 1971; Teis and Naidin, 1973; Riboulleau *et al.*, 1998; Rohl *et al.*, 2001; Gregory *et al.*, 2002; Poulsen *et al.*, in press; and others). According to these works with most complete data on the Oxfordian of Europe, paleotemperatures reflect primarily the latitudinal changes in this parameter. Temperatures fluctuations were substantial during the Jurassic, corresponding to three warming periods (Toarcian, middle Oxfordian, and Tithonian maximums) in Europe and Greenland (Bowen, 1969). The elucidated trend is complicated by insignificant deviations from average values in space and with time (early Kimmeridgian minimum). The asynchronous paleotemperature minimums and maximums in generalized curves compiled for various regions of the Boreal, Tethyan, and Notal provinces (Stevens, 1971; and others) created pessimistic attitude of some researchers to the absolute paleothermometry method (Hallam, 1978). In some works, there was admitted possibility of using oxygen isotopic ratios for paleotemperature reconstructions, although they are poorly correlative with data on carbon isotope ratios (Gruszczynski, 1998). Despite all the discrepancies in paleothermal studies, many of changes in absolute paleotemperatures estimated for Europe, general and particular, coincide however with changes in relative climatic characteristics based on distribution of corals (Arkell, 1933; Leinfelder, 1993; Bertling, 1993; Bertling and Insalaco, 1998; and others), dinoflagellates (Poulsen *et al.*, in press), and land flora (Vakhrameev, 1991; Abbink *et al.*, 1998, 2001; Rees *et al.*, unpublished data). The equator position proposed by Bowen is also compatible with that in the Arkell's reconstruction.

The available data on Jurassic paleotemperatures of Europe and adjacent regions are irregularly scattered. For instance, Bowen carried out his observations mainly in France and Germany and practically ignored other regions, except for England and Greenland. In terms of paleoclimatic approach, these data are sufficient to reveal succession of warming and cooling periods in different regions, although they give almost nothing for paleohydrological interpretation and can be even misleading in this aspect. It should be noted, that precisely paleoclimatic approach is a key objective of many interpretations.

In our country, paleotemperatures were estimated for the Russian platform and North Siberia (Saks, 1969; Teis and Naidin, 1973; and others). These observations within a wide stratigraphic interval are insufficiently accurate (stage intervals or greater). However, paleotemperature fluctuations defined in the Boreal Jurassic appeared to be synchronous with those in Europe. In the mentioned works, particular attention is paid to seasonal paleotemperature variations extremely valuable in terms of water mass characterization. The analysis of these observations showed in particular that the northern part of Eurasia was in the Late Jurassic under influence of water masses with temperatures of 10–14°C, i.e., transitional between subtropical and temperate. Even lower paleotemperature values (7 to 10°C) were obtained for high-boreal areas based on analysis of Middle–Upper Jurassic belemnites from Spitsbergen (Ditchfield, 1997).

Based on a brief review of principal results obtained by paleothermal studies of Jurassic marine sediments in northwestern Europe, it is possible to outline several characteristic aspects in approaches of different researchers to paleotemperature data and their interpretation. First, the studied materials (mainly belemnites) are sampled from a limited number of localities in the geographical region under consideration, and inferable trends are incomplete therefore, elucidating only the latitudinal variations of paleotemperatures. Second, the analyzed samples characterize discrete levels within a wide stratigraphic interval so that the maximal resolution rarely exceeds the stage rank. Paleotemperature data for narrower ranges (zone, subzone, and biohorizon) have been published not long ago only (Anderson *et al.*, 1994; Riboulleau *et al.*, 1998; Barskov and Kiyashko, 2000). Third, the paleotemperature curves compiled for long time intervals (e.g., Stevens' curves) characterize particular regions, e.g., the Russian platform or Europe, being considered, nevertheless, as global in significance. These curves completely ignore regional fluctuations, sometimes opposite in sign. Fourth, the results of paleotemperature analysis are usually interpreted as paleoclimatic changes, and spectrum of possible causal effects is depleted therefore, capable to lead in a dead end, because the water temperature is not a climatic parameter.

The traditional paleoclimatic approach to interpretation of paleotemperature data often ignores azonal character and factors of temperature distribution. Therefore, the paleotemperature fluctuations within a given latitudinal zone have been frequently interpreted as "anomalies" or artifacts related to technology of the analysis (Hallam, 1978). The probable influence of currents on temperature distribution is also rarely taken into consideration in reconstructions of the Jurassic seas paleotemperatures. An event, which is commonly under discussion, is the influence of the warm Tethyan current that presumably penetrated via Europe far northward. For example, Arkell assumed influence of a warm current during the Portlandian in England based on coexistence of thermophilic marine and cryophilic continental faunas (Arkell, 1961). Bowen (1969), who noted strong paleotemperature variations (up to 6°C in range within area 200 km across) in the Parisian basin located slightly southward, explained this phenomenon by a local warm current. For the Late Jurassic, he also assumed influence of a warm current that crossed the North Sea and reached Greenland. This event was also discussed in reconstructions based on fossil floras (Abbink *et al.*, 2001). The origin of all these currents

has been considered in the work by Teis and Naidin (1973) who attributed them, based on additional data from Siberia, to a single warm current flowing from southern Europe via Greenland to Siberia. However, the general water circulation system in Jurassic basins has not been reconstructed because of fragmentary character of data used.

The paleobiogeographic data, if they are casually connected with paleotemperatures, are more informative than absolute temperature and better elucidate the hydrological regime in the given paleobasin estimates, being preferable because they are already available and need in a comprehensive analysis only. There are many works dedicated to this problem, and reconstructions of paleocurrents based on distribution of various faunal groups, e.g., ammonites, are of particular interest among them. In terms of methodology, these reconstructions are based on the assumption that migration paths of ammonites coincided with and were controlled by currents, being oriented away from the evolution center of this faunal group (Khudolei, 1976). The countercurrents in the strait between Greenland and Scandinavia (Enay, 1980; Khudolei, 1984, 1985, 1988, 1993; Abbink *et al.*, 2001) have been assumed for Jurassic seas in addition to the cold water discharge in the Pechora basin (Enay, 1980) and cold current of the Callovian time from the Central Russian sea to the Central Asian basin (Khudolei, 1985).

In this work, the main attention is focused on reconstruction of a general circulation system for the Callovian–Oxfordian seas of northwestern Eurasia based on ammonites. In my previous work (Kiselev, 1998), I considered the relative paleotemperatures only for the Central Russian sea. Main conclusions of that work were as follows. (1) In ammonite assemblages from two and more sections, proportions between taxa of different origin change in meridional direction even within geographically limited areas, i.e., these changes are consistent with the zonality principle. (2) Ammonites are mostly stenothermal organisms highly sensitive to water temperature gradients, and paleobiogeographic data on their distribution can be consequently used for detailed reconstructions of the thermal regime in sea basins. (3) Secular changes in relative paleotemperatures vary in different regions (Central Russian sea and adjacent areas of the Tethys, Europe, and the Arctic basin) and even subregions (western and eastern zones of the Central Russian sea), thus being dependent on regional circulation systems. In the Central Russian sea of the Callovian–Oxfordian time, there was a steady circulation system that included a permanent countercurrent crossing the entire basin from the north southward and several warm countercurrents of different origin.

METHODS

In this work, I used principle of the relative paleotemperature analysis applied to ammonite assemblages. This method of estimating relative paleotemperatures has been described earlier (Kiselev, 1998). Four assumptions are basic in the method.

(1) The thermal factor had decisive impact on geographic distribution of ammonite *species* within *intercommunicating epicontinental seas.* In oceanic basins, ammonites belonged to the other biome (in terms of Thierry, 1988), and their distribution was controlled therefore by shelf boundaries and by other factors of morphological–functional adaptation (Dommergues *et al.*, 1987; Thierry, 1988).

(2) The probability of species origin in a diversification center of a genus is maximal. The corresponding center coincides with the species radiation center. Migration paths are oriented away from the species radiation center (Khudolei, 1976).

(3) Four main centers of species radiation were characteristic of Jurassic seas: (a) the Arctic basin (Boreal Province, after Zeiss, 1968), (b) the Central Russian sea (Subboreal Province, ibid), (c) the European basin (Subboreal and Submediterranean provinces, after Ziegler, 1963), and (d) the Tethyan basin (Mediterranean and southerly provinces, ibid). The belonging of a species to a certain radiation center is determined by the assumption (2) regardless of its distribution. For widespread forms, the probability of origin in a radiation center is always maximal, though lesser than 1. An endemic species only can be considered as having probability of origin in a particular basin that is equal to 1 (until the established occurrence in the other basin).

(4) Each of four radiation centers was located at different paleolatitudes, in the following N–S succession: Arctic basin, Central Russian sea, European basin, Tethys ocean. This assumption leans upon paleogeographic reconstructions, the mobilistic included (Cariou *et al.*, 1985; Dommergues *et al.*, 1987; Golonka, 2002; Golonka and Bocharova, 2000; and others). In line with the principle of climatic zonality, the average temperature of water masses increased in the same direction.

The technical aspect of the method consists in calculation of exact proportions between species of different origin in each ammonite assemblage. Inasmuch as the origination center of any species, in particular cosmopolitan one, cannot be determined, in the absolute sense, the *probability* of its origin is calculated for each faunal province using the formula:

$$P_{\text{species origin}} = \frac{q}{\Sigma q}, \quad (1)$$

where q is a sum of species belonging to the particular genus in the relevant basin, Σq is a sum of q in all basins where the species occurs. The formula of species origin probability is based on the assumption (2). Four or less P values are calculated for each species depending on a number of basins, where it was found. For example,

four P values calculated for *Cardioceras cordatum* (Sow.), the cosmopolitan species within northwestern Eurasia, are 0.1, 0.44, 0.21, and 0.25 for the Tethyan, European, Central Russian, and Arctic basins, respectively. When the species is not cosmopolitan, three or less P values are calculated. For endemic species, P is equal to 1.

When calculating proportions of species originated from different radiation centers in a given ammonite assemblage, all P values are summed with subsequent averaging. In order to obtain parameters characterizing paleotemperatures, conventional coefficients 1, 2, 3, and 4 in order of the average water temperature rise are introduced into the next formula according to the above assumption (4):

$$\mathrm{RPC} = \frac{4\Sigma\mathrm{Pt} + 3\Sigma\mathrm{Pe} + 2\Sigma\mathrm{Pc} + 1\Sigma\mathrm{Pa}}{\Sigma\mathrm{Pt} + \Sigma\mathrm{Pe} + \Sigma\mathrm{Pc} + \Sigma\mathrm{Pa}}, \quad (2)$$

where RPC is the relative paleotemperature coefficient, Pt, Pe, Pc, and Pa are probabilities of species originated in the Tethyan, European, Central Russian, and Arctic basins, respectively. Inasmuch as the probability of species origin for the entire system of basins is equal to 1, denominator $\Sigma(\Sigma P)$ in the formula (1) is consequently equal to the number of species in each assemblage, and the formula turns into the following:

$$\mathrm{RPC} = \frac{\Sigma(\mathrm{Ca, c, e, t}\Sigma\mathrm{Pa, c, e, t})}{N}, \quad (3)$$

where C is a conventional coefficient characterizing the average water temperature rise in different basins, and N is number of species in a given ammonite assemblage. Since the RPC is causally connected with variable temperature regimes in the system of basins under consideration according to assumptions (1) and (4), it can be referred to as "relative paleotemperature."

There are two stages of RPC calculation. The first stage includes compilation of a complete list of species occurring in all the basins within the particular stratigraphic interval and calculation of q, Σq, and P parameters for each genus separately. The second stage consists in the RPC calculation for the all species of different genera from the particular stratigraphic unit of the individual section or locality.

At the next stage, the calculated RPC values are to be plotted and outlined by isolines on the map. In total, six paleotemperature isoline schemes have been compiled for six time equivalents of the Callovian and Oxfordian substages. To fulfill this task, the necessary parameters have been calculated for 205 ammonite assemblages from 361 localities in 54 areas of the Arctic, European (Peri-Tethyan), Tethyan, and Central Russian basins of northwestern Eurasia (Table 1). The source materials have been collected from publications, museum and private collections in addition to original data obtained during field works in the Russian platform.

Despite the relative simplicity of this method, its application is associated with the following disadvantages.

(1) *Incompleteness of stratigraphic and paleontological data.* Ideally, the RPC should be calculated for the each outcrop, but a study area can frequently be characterized by a single site of several outcrops only because of incompleteness of other sections and fossil records (as is known, compilation of the complete list can take tens of years even in case of regular observations). Because of the same reason, the RPC calculation is problematic for narrow stratigraphic intervals (a zone and lesser subdivisions) in large territories, and therefore the substage scale is used in this work. By a general characterization of the substage time span, which are 1.5 to 2.0 m.y. long in the Jurassic System (Callomon, 1984), many fluctuations in the thermal regime of sea basins, which undoubtedly took place because of cyclic processes in the hydrosphere, remain obscure. Accordingly, any reconstruction for such a time interval depicts only most general trends the thermal regime evolution in a basin. In principle, the paleotemperature parameters and their changes can be inferred for any time interval, if configuration and disposition of the basins did not change. Within the greater time intervals, the calculated parameters become averaged, but their spatial distribution remains unchanged.

(2) *Subjective systematics.* As is known, researchers have different, even opposite views on taxonomy of fossils. Lists of taxa compiled by adherents of species unification and sexual dimorphism concept will be shorter than those of other paleontologists will, and calculated RPC values should be different therefore. Nevertheless, my own experience shows that different taxonomic approaches create an insignificant discrepancy. The erroneously identified close and endemic species almost do not affect the RPC value, because they occur usually in a single paleobiogeographic province (for endemic species this value is always equal to 1). For example, P_{origin} for *Cadoceras elatmae* (Nik.) in the Arctic, Central Russian, European, and Tethyan basins is 0.63, 0.25, 0.11 and 0.01, respectively. For *C. septentrionale* and *C. emelianzevi* Vor., the values are identical, equal to 1, 0, 0, 0 as both forms are unknown beyond the Boreal Province. The error in identification of close species with different geographic distribution is also insignificant, as it is usually concealed under the highest P value characterizing the province (basin), where the genus diversification center was. *Cadoceras tschernischevi* Sok. is known from the Central Russian and Arctic basins, therefore its P_{origin} is 0.69, 0.31, 0, and 0, i.e., insignificantly different in general as compared to above values calculated for *Cadoceras elatmae* (Nik.). The experience shows that the RPC value changes insignificantly and does not affect the general interpretation, if the sum of errors is 20% or less.

(3) *Taxonomic diversity increase.* The discovery of new species increases their lists for any locality. This should change unavoidably the RPC values, particularly when these species represent taxa untypical of the given biochore. Par example, the Callovian and Oxfordian families Oppelidae, Pachyceratide, and Reineckeidae represent taxa of this kind for the Boreal Province and the family Cardioceratidae for the Tethyan Realm. Therefore, it is necessary to estimate how sensitive is the method and how persistent will be conclusions after discovery of additional ammonite taxa, especially in poorly studied areas. I tested the issues using the cases of most critical increase in diversity of ammonite species from alternative biochores (Fig. 1). The RPC value depending on the species diversity changes as the power function, i.e., the rate of changes gradually decreases to reach a plateau. It is remarkable that the RPC change is insignificant even by substantial growth of the species list. The twofold increase in species number gives the enlargement factor of 1.35 only for the RPC parameter. Such an increase of species diversity is unreal for in any alternative biochores, particularly for those hosting tens of species (e.g., for well-studied Europe). Otherwise, it would be unreasonable to discriminate biochores. Most probable is the species diversity (species list) increase created by species of local origin or from a neighboring biochore. In this case, the rate of RPC changes is substantially lower, unaffecting the results. Thus, the available data on diversity of ammonites in the Callovian and Oxfordian sediments should be considered as representative and sufficient for solving the posed problems.

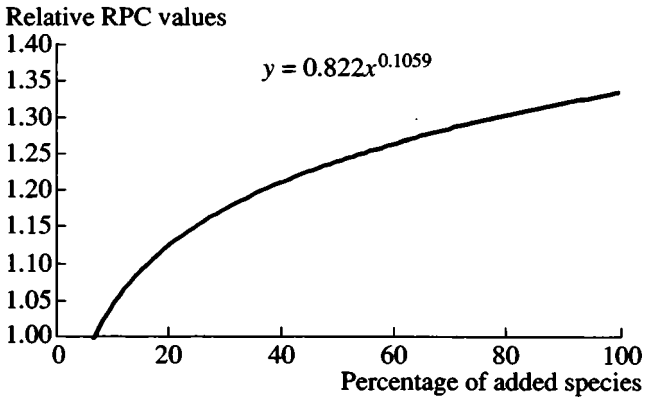

Fig. 1. Changes of relative paleotemperature coefficients (RPC) as depending on species diversity increase in ammonite assemblage from alternative biochores (explanation in the text).

RESULTS

RPC changes in space. The distribution of RPC values over the territory of northwestern Eurasia can be presented in a form of paleotemperature charts of relative paleoisotherms (RPI), which illustrate zonal and regional aspects of the thermal regime evolution in the Callovian–Oxfordian seas (Figs. 2–7).

The azonal (regional) distribution of RPC values is detectable in isoline charts from bends of latitudinal and sublatitudinal isoline trends. In maps of present-day water temperatures, such bends outline the currents (Lednev and Muromtsev, 1953). The bend orientation southward and northward characterizes cold and warm currents, respectively, in the Northern Hemisphere. Therefore, despite their relative character and hypothetical relation with paleotemperature isolines, the observable RPI bends are interpreted here as hypothetical markers of paleocurrents.

The reconstructed paleocurrents characterize the general circulation system in basins of northwestern Eurasia and its evolution during the Callovian–Oxfordian.

For the *early Callovian* (Figs. 2, 8), there are traceable both the cold and warm currents. Two cold currents are distinguishable to the west and east off the Scandinavian massif. The eastern current penetrates into the Central Russian sea via the Pechora seaway. Previously (Kiselev, 1998), I regarded it as an effluent from northerly parts of the Boreal basin. In the Pechora basin or Middle Volga region, this current bifurcates into western and eastern branches: the Dnieper and Caspian seaways, respectively. The Dnieper cold current probably reached, via the Orekhovskii strait, the Crimean Tethys (the term of Sazonova and Sazonov, 1967), where it likely bifurcated again. According to mentioned authors, the Orekhovskii strait did not exist in the Callovian, although a significant percentage of Boreal forms in the Crimean ammonite assemblage and related RPI bend lead to an opposite conclusion. The eastern branch in the Caspian region probably flowed to Central Asia (Pamir Province, after Amanniyazov, 1971).

Another cold current bypassing the Scandinavian massif on the north was likely an effluent stream as well. In the north, it approached Greenland and then deviated eastward penetrating into Europe between England and Norway. Its fairway crossed southern Germany, where boreal species are particularly abundant, and then continued to the Tethys. Both cold currents served probably as main migration paths for boreal ammonites. This assumption is consistent with conclusions of Thierry (1976) who defined two migration paths for boreal ammonites in the Callovian.

Opposite northward-oriented bends of isolines outline warm currents in the European and Central Russian seas. The European warm current began in the Tethys and initially flowed to the Polish sea, being probably responsible for a wide development of carbonate facies there. Farther northward, the current should unavoidably deviate because of the Scandinavian massif, washed the latter on the south and flowed via the Pripyat strait to the Dnieper basin of the Central Russian sea. From the latter, the warm current followed along the Scandinavian massif up to the Pechora sea-

Table 1. Relative paleotemperature coefficients for 54 areas of northwestern Europe[1]

Area	NL	Callovian						Oxfordian					
		early		middle		late		early		middle		late	
		RPC	*n*	RPC	*n*	RPC	*n*	RPC	*n*	RPC	*n*	RPC	*n*
Scotland (Sky) P	5	–	–	–	–	–	–	–	–	2.56	18	2.22	22
England P	25	2.92	70	3.13	39	2.92	60	2.91	74	2.89	72	2.71	44
Germany (Swabia) P	6	3.06	93	3.27	61	3.11	33	3.02	47	2.9	56	2.91	36
Germany (Franconia) P	7	3.2	39	3.13	36	3.04	36	–	–	3.15	54	3.06	84
Northern Germany P	6	3.11	21	3.18	38	3.04	45	2.84	24	–	–	–	–
Poland, P, C	15	3.22	35	3.38	31	2.95	25	2.98	54	2.95	27	3.04	64
Spain (Pyrenees) P	2	3.46	19	3.79	10	–	–	3.66	8	3.49	24	3.54	28
Spain (Ricla) P	4	3.82	44	3.91	43	3.95	9	–	–	–	–	–	–
Southeastern Spain P	3	–	–	–	–	–	–	3.6	8	3.44	19	3.65	38
Portugal (Cap-Mondego) P	6	3.81	59	3.89	14	3.49	8	–	–	–	–	–	–
Central France P	8	3.29	41	3.38	61	3.15	114	2.79	56	–		2.83	9
Eastern France P	19	–	–	3.38	61	2.77	37	–	–	2.8	26	2.99	58
Southern France (Maritime Alps) P	3	3.62	21	3.73	25	–	–	–	–	–	–	–	–
Normandy P	8	–	–	–	–	–	–	2.68	13	2.85	12	2.71	8
Northern Jura (Herziah) P	1	–	–	3.25	40	3.08	62	3.11	56	3.15	23	3.05	52
Southern and Middle Jura P	7	3.44	8	–	–	3.37	35	2.91	33	3.05	84	3.1	66
Italy P	4	3.44	13	–	–	–	–	–	–	3.41	12	3.77	13
Sicily P	2	–	–	3.8	9	–	–	3.65	11	–	–	–	–
Moravia P	1	–	–	–	–	–	–	–	–	3.13	23	–	–
Romanian Carpathians P	2	3.64	13	3.85	9	–	–	3.36	6	3.51	6	3.65	10
Dniester, Prut P	5	–	–	–	–	3.35	5	–	–	–	–	3.77	9
Baltic region P	7	–	–	3.18	25	2.93	34	2.65	7	2.7	24	2.3	29
Pomor'e P	4	–	–	–	–	3.04	11	2.93	17	2.88	14	2.66	18
Donets Jurassic P,C	8	2.78	26	3.13	9	2.86	18	2.88	24	2.91	19	2.47	5
Zheleznogorsk C	1	2.32	13	2.92	8	–	–	–	–	–	–	–	–
Staryi Oskol C	1	–	–	–	–	–	–	–	–	2.63	20	–	–
Moscow region P, C, O	14	–	–	2.97	13	2.67	11	–	–	–	–	2.01	12
Rybinsk area O, P, C	9	2.26	7	2.5	16	2.62	15	2.59	12	2.4	26	2.19	12
Nizhnii Novgorod region (Prosek) O, C, P	3	2.02	31	3.1	–	–	–	–	–	–	–	2.1	10

Table 1. (Contd.)

Area	NL	Callovian						Oxfordian					
		early		middle		late		early		middle		late	
		RPC	*n*	RPC	*n*	RPC	*n*	RPC	*n*	RPC	*n*	RPC	*n*
Nizhni Novgorod region (Uzhovka) O, P	2	2.26	13	2.93	23	–	–	–	–	–	–	–	–
Oka River (Alpat'evo–Elat'ma) O, C, P	11	2.36	15	2.64	72	2.89	36	2.57	27	2.52	34	2.02	14
Mikhailov O, C	2	–	–	2.85	47	2.76	29	2.71	9	–	–	–	–
Unzha O, C, P	17	2.14	22	2.27	37	2.38	15	2.24	–	2.48	–	2.12	–
Sysola O, C, P	3	2.21	10	2.15	24	–	–	–	–	–	–	–	–
Orenburg P, C	4	–	–	2.92	7	2.56	13	2.52	12	2.33	11	2.13	7
Saratov Volga region P, C	8	2.47	16	2.75	22	2.76	55	2.57	27	2.69	28	–	–
Pechora River basin P, C, O	13	1.58	12	2.25	8	1.71	15	2.04	3	2.2	5	2	15
Eastern Greenland P	22	1.93	24	2.52	4	1.81	10	2.33	8	2.24	10	1.96	19
Novaya Zemlya P	4	–	–	–	–	1.78	8	–	–	–	–	–	–
Spitsbergen P	9	–	–	–	–	1.59	5	2.05	6	–	–	1.92	9
Mangyshlak P, C	6	3.34	28	3.29	21	3.26	26	2.88	18	–	–	2.68	8
Tuarkyr P	12	3.78	38	3.6	67	3.49	95	3.24	20	3.02	15	3.63	8
Kugitang-Tau P	6	3.69	30	3.66	30	3.58	32	3.09	14	3.55	14	–	–
Crimea P, C	2	3.27	13	3.63	7	3.42	10	3.33	10	3.32	4	3.53	6
Northern Caucasus P	14	3.49	69	3.62	91	3.45	55	3.25	12	3.22	9	3.85	11
Southern Caucasus	10	3.66	28	3.56	17	3.56	6	–	–	3.56	11	–	–
Negev P	3	–	–	3.91	13	3.79	18	–	–	–	–	–	–
Lebanon–Antilebanon P	5	–	–	3.72	7	3.72	4	3.82	41	3.63	4	–	–
Iran P	2	–	–	3.83	31	–	–	–	–	–	–	–	–
Kopetdag P	2	–	–	–	–	–	–	–	–	3.67	9	3.6	11
Western Gissar P	3	–	–	–	–	–	–	3.29	7	–	–	–	–
Morocco P	2	–	–	–	–	–	–	3.51	8	–	–	3.6	6
Algeria–Tunisia P	5	3.73	17	3.8	43	3.45	7	–	–	3.69	18	–	–
East Africa P	8	–	–	3.95	18	3.81	13	–	–	3.88	10	3.89	15

[1] Letters designate sources of data used for compilation of ammonite species lists: (P) published (works are not listed because of their abundance); (C) collections; (O) original data; (NL) number of localities, which were used for compiling integral faunal lists for each region and for which relative paleotemperature coefficients (RPC) were calculated; (n) number of species in the ammonite assemblage. Dash indicates the absence of representative data for RPC calculation for the given time interval.

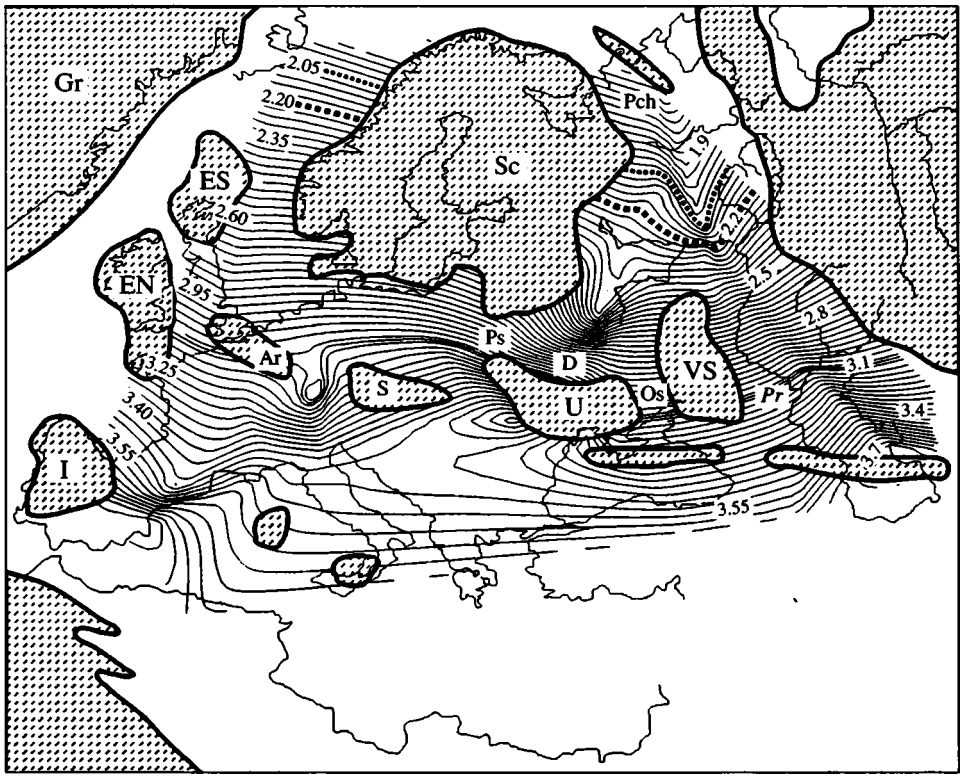

Fig. 2. Relative paleoisotherms (RPI), the early Callovian of northwestern Eurasia (processed data are from Sazonova and Sazonov, 1967; Thierry 1976, 1982; Cariou *et al.*, 1985; Dommergues *et al.*, 1987a, 1987b; Ziegler, 1990; Stampfli *et al.*, 1998). Geomorphological features: (EN) England–Normandy land; (Ar) Ardennes massif; (ES) England–Scotland land; (VS) Voronezh–Stavrolpol land; (Gr) Greenland; (I) Iberian massif; (Os) Orekhovskii strait; (Pr) Pripyat strait; (D) Dnieper basin; (NC) North Caspian basin; (Pch) Pechora basin; (S) Sudeten massif; (Sc) Scandinavian massif; (U) Ukrainian massif. In this and other charts, isolines are plotted by the Natural Neighbor triangulation method using the program Surfer 7. Dotted isoline 2.05 approximately corresponds to the thermal boundary between tropical and subtropical water masses (see text); thick isoline 2.2 approximates the thermal distribution boundary of the present-day of hermatypic corals (explanation in text).

way. This inference is supported by numerous finds of European faunal elements in the Dnieper basin, which implies the unidirectional migration via the Pripyat seaway already in the Callovian (Yanin, 1998, 1999). The warm current could be a decisive migratory factor.

In the *middle Callovian*, the RPI shift is oriented northward in all basins. This is indicative of substantial water warming in the entire region, which promoted the northward migration of thermophilic marine biota, the ammonites and other groups of marine invertebrates included (Yanin, 1998). The general circulation system was presumably the same as in the early Callovian, consisting of two cold effluent currents, the S–N trending European warm current, and the Dnieper warm countercurrent, the orientation and disposition of which remained unchanged (Figs. 3, 8). Simultaneously the system included some new elements. For example, there are several west-oriented RPI bends in the Tethys, which probably outline a latitudinal current that encircled the planet along the equator carrying waters westward (Hallam, 1978); Enay, 1980; Westermann, 1981). Its present day analogue, the north equatorial (trade-wind) current, is located 5 to 10° (in equinox periods) northward of the equator (Lednev and Muromtsev, 1953). Judging from the RPI bends, the presumable current turned northward at the latitude of Romania to meet the Ukrainian land, where it bifurcated into the main branch forwarded northward to the Polish sea and the other one that flowed along eastern coast of the Ukrainian land to meet the cold current from the Central Russian sea.

In the Tethys, there is distinguishable a presumable trace of another east-oriented warm current. According to the RPI bends, it passed between the Spanish (Iberian) archipelago and North Africa almost parallel to the above-mentioned equatorial current. The assumed current could reach Tethys via a long narrow strait (Hispanic Corridor, after Westermann, 1984, 1993) from the Caribbean basin, where there could (and most likely was) its own circulation system connected with additional water sources and specific discharge ways. Therefore, it could be an effluent stream from the Caribbean basin. Topologically, the current in question resembled the Gulf Stream, although it was not obligatory the genetic predecessor of the latter. The Gulf Stream is fed by autonomous trade-wind currents,

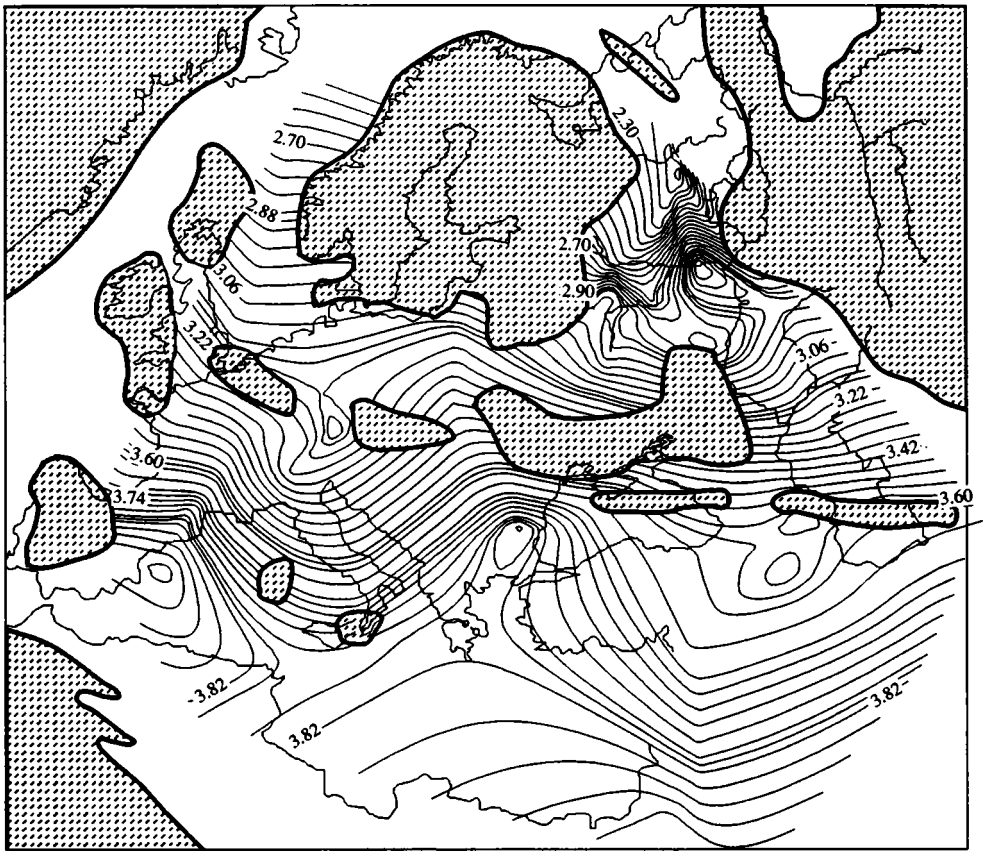

Fig. 3. Relative paleoisotherms (RPI), the middle Callovian of northwestern Eurasia.

being a warm current of the oceanic type, while the Hispanic Corridor, the early prototype of the Atlantic Ocean ("epicontinental Proto-Atlantic" after Thierry, 1982) was not an ocean in hydrological terms. The corridor was floored in the Callovian mainly by the continental crust (Westermann, 1981, 1993). Most likely, the proto-Caribbean Sea represented a large evaporation basin, as it follows from abundance of evaporites in that region (Hallam, 1978). The basin contained very warm hypersaline and heavy waters flowing to the Tethys via the Hispanic Corridor. There are grounds to believe that the system of Tethyan currents was of a similar type in the early Callovian as well (this is evident from the similar disposition of non-Tethyan currents), although it is impossible to test this assumption because of incomplete data on early Callovian ammonite assemblages of the Tethys.

In the middle Callovian, the circulation system of the Central Russian sea was slightly more intricate than that of the early Callovian time. The Boreal effluent current did not reach anymore the Tethys via the Crimea (Orekhovskii strait became closed, Sazonova and Sazonov, 1967). It met instead the Dnieper warm current that deviated as if under pressure of cold waters, as one can judge from the behavior of isotherms. The Dnieper current flowed, as in the early Callovian, along the Scandinavian massif, where the convergence zone of warm and cold waters could be formed. In this area, the cold current sharply curved around the Voronezh paleoland and then flowed into the Tethys via the North Caspian basin. In the Tethys, the Central Russian cold current turned westward and met the warm Tethyan current.

In the *Late Callovian*, isolines in the northern part of the region are displaced southward to the latitude of England (to the RPI 2.95 in Fig. 4), while southerly isotherms remain in the same position as before. This change might reflect cooling northward of the England–Scotland massif. The total circulation system remained approximately unchanged, and all main currents occupied the same position (Fig. 8). Small transformations are as follows. (1) West Tethyan current penetrated the northern part of the European sea (north of the Ardennes massif) beginning from Germany, not from the Polish sea. It cannot be ruled out that this current behaved similarly in the early Callovian as well, although this is not reflected in isotherms. (2) The western branch of Boreal current in the Central Russian sea entered the Dnieper strait, followed slightly southward parallel to the warm current, and then flowed to the European basin via the Pripyat strait. These currents probably flowed at different depths. (3) Judging from

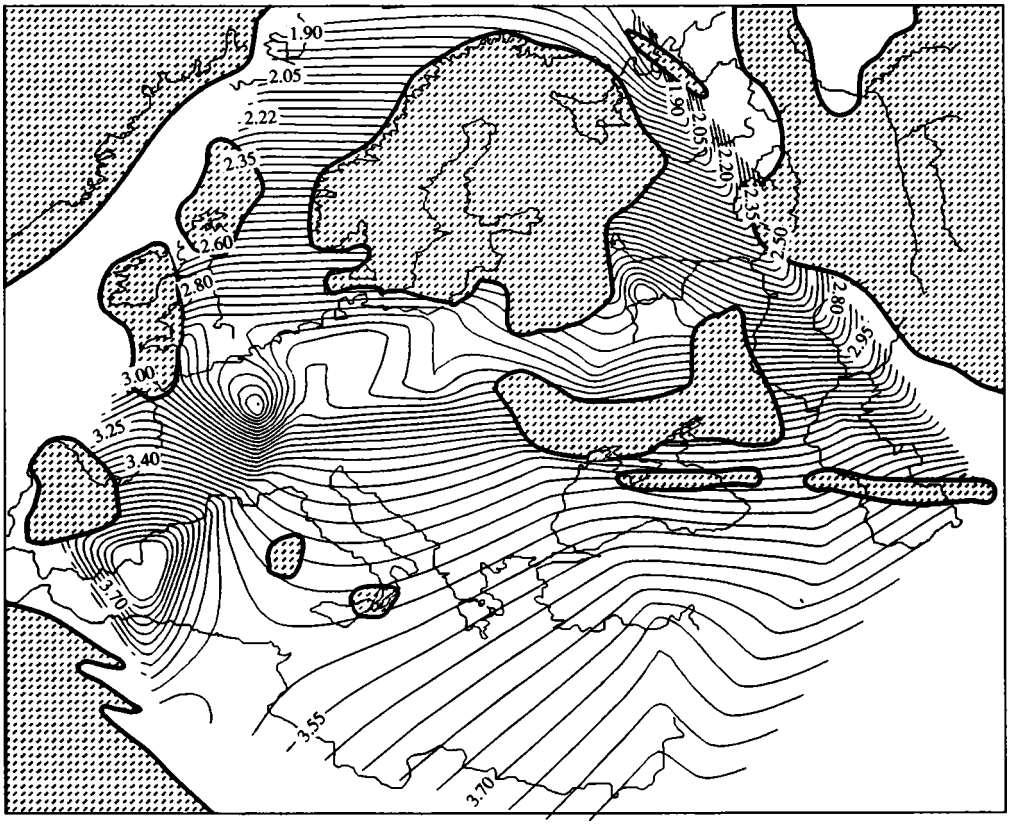

Fig. 4. Relative paleoisotherms (RPI), the late Callovian of northwestern Eurasia.

isotherms, the Central Russian convergence zone disappeared.

In the *early Oxfordian*, the presumable configuration of warm and cold currents did not change in general (Figs. 5, 8). In distinction from the middle Callovian epoch, the west effluent current of Boreal origin crossed Normandy, not Germany, and should pass west of the British Islands, although this is not reflected in isotherms. Simultaneously, the western Tethyan current avoided the Polish sea and flowed toward the England–Normandy massif. There, the warm and cold currents joined to form the convergence zone that is outlined in isotherms. The Central Russian sea received a lesser quantity of European waters than in the late Callovian epoch, being dominated by Boreal fauna that migrated via cold effluent currents.

In the *middle Oxfordian*, the system of detectable currents was the same in general as before (Figs. 6, 8). On global scale, the RPI contours are slightly displaced northward almost everywhere that may be indicative of warming in the entire region. This probably was a consequence of intensified Tethyan currents, since the northward-oriented RPI bends became more convex, particularly in the Central Russian sea.

In the *late Oxfordian* chart, isotherms are sharply displaced southward (Fig. 7), and a new RPI set corresponding to low RPC values appeared. This is particularly well noticeable in areas northward of England and Normandy, like in the late Callovian. This shift likely reflects cooling and ingression of Boreal waters that was of a much higher amplitude than in previous epochs. The ingression was particularly significant in the Central Russian sea that was entirely occupied by Boreal waters (see RPI marked by low values in charts).

The general circulation system in the northwestern Eurasian seas retained its principal features, although some details changed (Fig. 8). The Central Russian sea was probably beyond the influence of warm Tethyan currents, which was characteristic of the middle Oxfordian. The Boreal effluent current significantly intensified was responsible for cooling in the entire Central Russian sea, the Dnieper basin and adjacent Tethyan areas (Mangyshlak, Kopetdag) included. The Dnieper basin was again influenced by the Boreal current at the time of the Orekhovskii strait wide opening (Sazonov and Sazonova, 1967). In the northern part of the Dnieper seaway, there was a convergence zone of warm and cold waters resembling that of the middle Callovian time. However, boreal waters more intensely invaded the Dnieper seaway west of the Voronezh land (Ryazan region), from where they flowed to the Orekhovskii strait. The Dnieper warm current moving northward met the Boreal countercurrent and deviated partly southward. As a result, a circular warm current

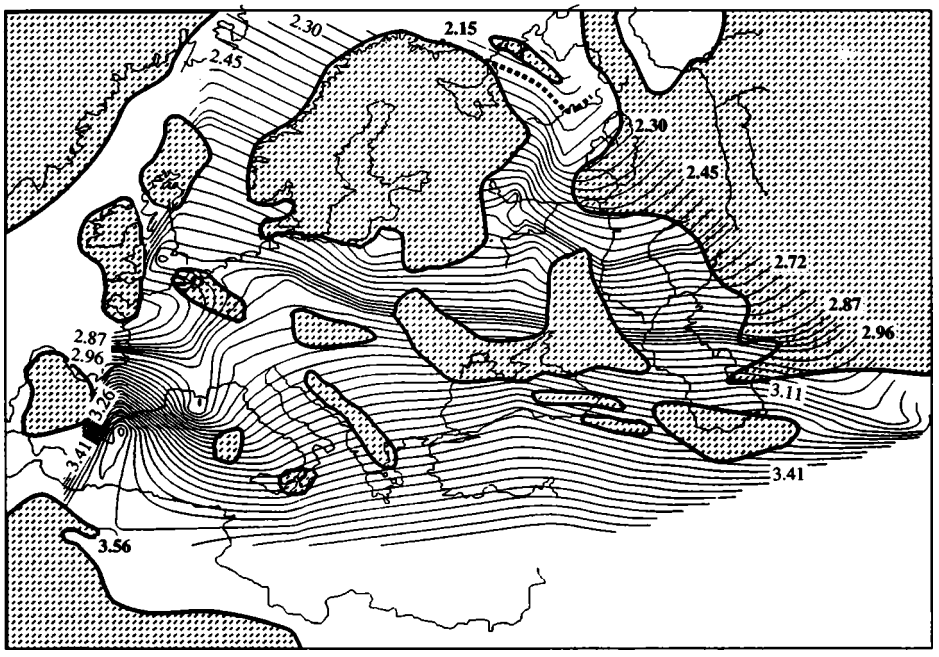

Fig. 5. Relative paleoisotherms (RPI), the early Oxfordian of northwestern Eurasia.

originated in the internal part of Central Russian sea during the late Oxfordian, like in the middle Callovian. Its influence is confirmed by finds of scleractinian corals in the axial part of "warm" isotherm bends (Fig. 10).

The cold current flowing out of the Orekhovskii strait met a powerful warm current from the Tethys in the chart area marked by blunt bends of "warm" isotherms and by wedging-in "cold" isotherms. Simultaneously, the higher density of isotherms is indicative of a convergence zone in that area.

The RPI zoning is distinct in all the charts displaying the regular decrease of RPC values from the south northward. This is a manifestation of the water cooling trend toward higher latitudes. According to widely accepted views, the Jurassic latitudes differed from the modern ones. The equator was located at that time approximately 30° northward of its present-day position and crossed the southern Mediterranean basin (Arkell, 1956; Bowen, 1969; Thierry, 1976; Cariou *et al.*, 1985; Dommergues *et al.*, 1987; Stampfli *et al.*, 1998; and others). This is substantiated by absolute and relative paleotemperature data that characterize equatorial water masses.

The annual temperature variations represent an important characteristic of water masses. The amplitude of temperature fluctuations can also characterize the substantially longer time intervals, if positions of continents relative to the equator changed insignificantly. Moreover, they characterizes also the thermal zoning of a given water mass, as it is shown in previous work (Kiselev, 1998). Variations of the RPC values throughout the Callovian–Oxfordian period (Fig. 9) show the regular dependence of their amplitude on the latitude, though with some regional deviations. In the southernmost part of the region under consideration corresponding to the northern Tethys, the RPC variations are almost negligible. The lowest temperature fluctuations in modern oceans are characteristic mainly of tropical water masses, where they are as low as 1 to 3°C (Zhizhchenko, 1974), corresponding to 3–12% of average annual variations. Almost all over the region under consideration, the RPI variations do not exceed 10%, and water basins were situated therefore in tropical zone, the northern boundary of which was probably located slightly northward of the Scandinavian massif, although its position undoubtedly migrated concurrently with changes in hydrological regime.

As was noted, the absolute paleotemperatures are calculated for a limited number of geographic sites and stratigraphic levels. In such a situation, it is impossible to reconstruct an adequate evolution of thermal regime in the basins. Nevertheless, extrapolating the absolute data to the relative ones we can, though with reservations, to draw absolute isotherms through different-size area based on calculated relative values. The calculated paleotemperatures of the late Oxfordian are published for eastern Greenland, France, and Switzerland (Bowen, 1969; Padden, in press), Scotland (Tan *et al.*, 1970), Spitsbergen (Ditchfield, 1997), and European Russia (Riboulleau *et al.*, 1998). Being compared with the late Oxfordian RPI chart, they elucidate some aspects of the generalized model discussed above. With due regard for discrepancy between data obtained by different methods of paleothermometry and for doubtful information on many time intervals, I consider here

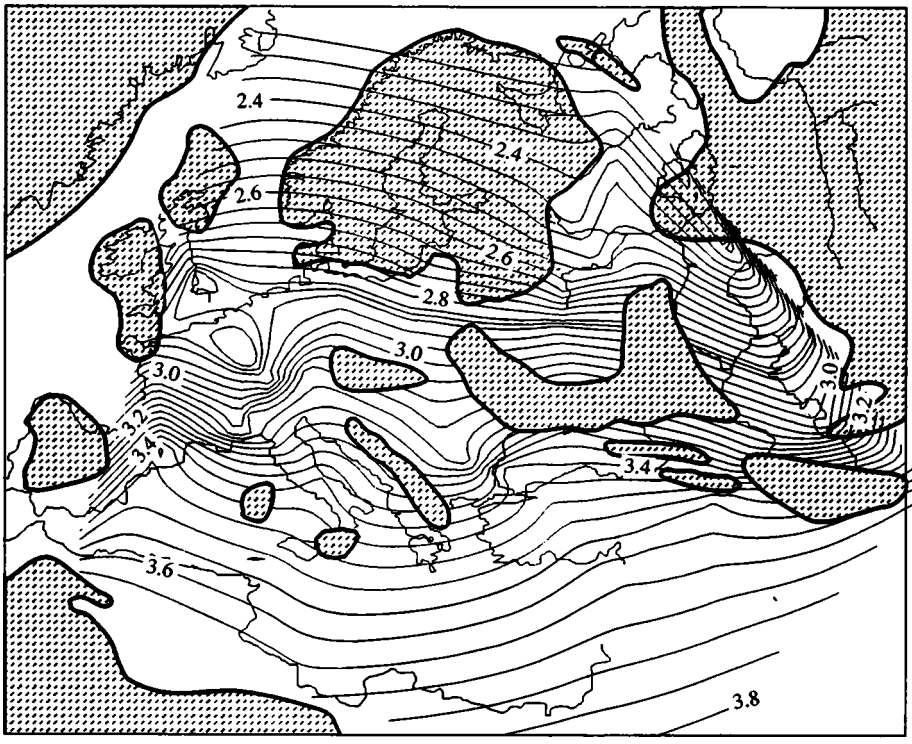

Fig. 6. Relative paleoisotherms (RPI), the middle Oxfordian of northwestern Eurasia.

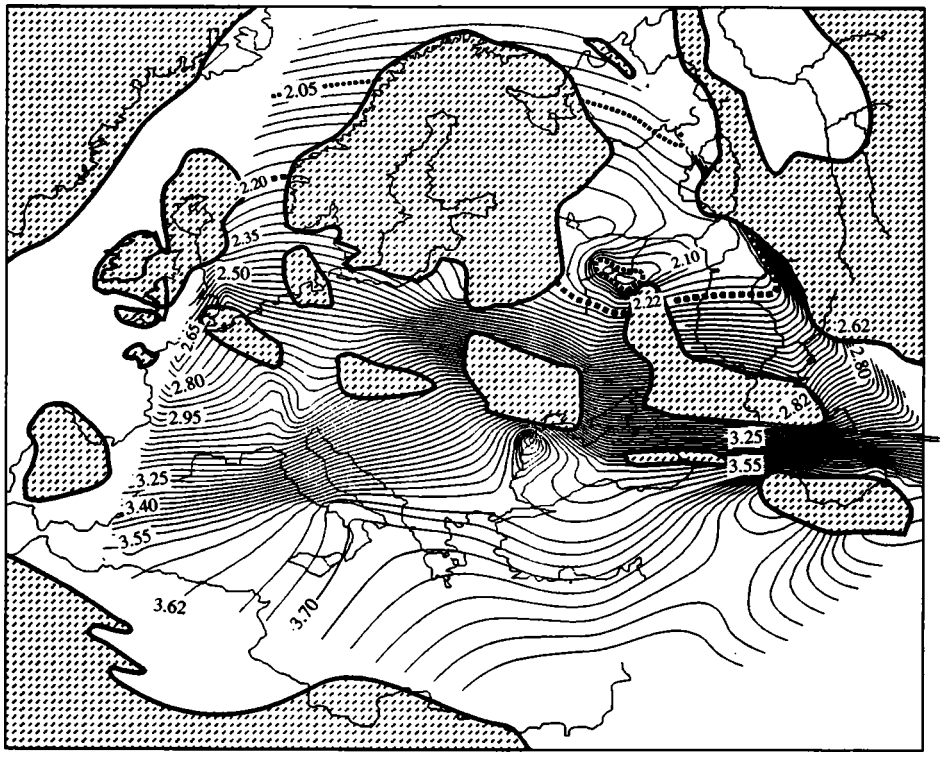

Fig. 7. Relative paleoisotherms (RPI), the late Oxfordian of northwestern Eurasia.

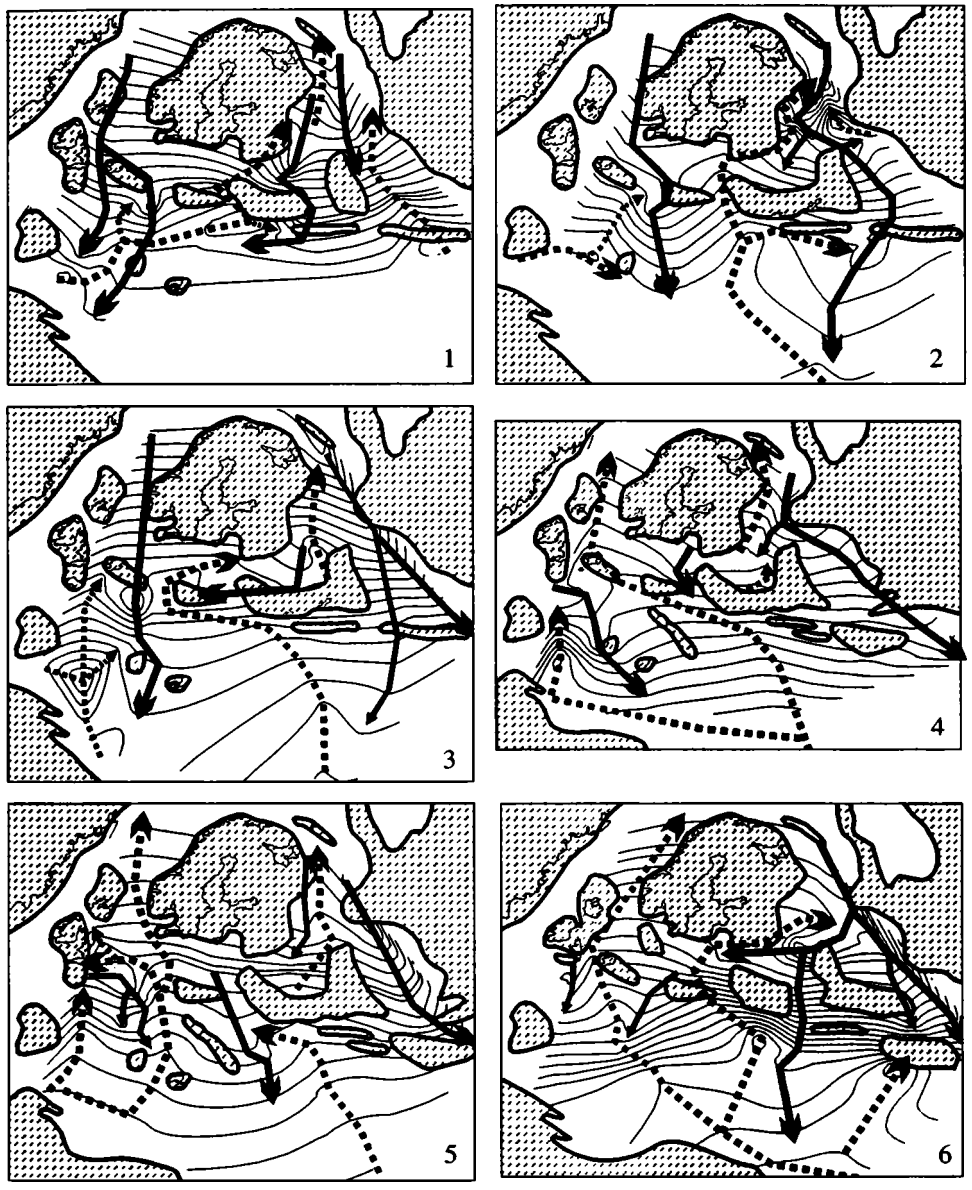

Fig. 8. Reconstructed circulation systems in Callovian–Oxfordian seas of northwestern Eurasia; solid and dotted arrows connecting peaks of RPI bends show presumable directions of cold and warm currents. Because of deficient information on some areas and wide extrapolations, the arrow lines locally cross the land blocks unlike the real currents: (1) early Callovian; (2) middle Callovian; (3) late Callovian; (4) early Oxfordian; (5) middle Oxfordian; (6) late Oxfordian.

only a chart for the late Oxfordian (Fig. 10). The chart is based on the absolute paleotemperature analogues (APA) extrapolated to RPC values of the late Oxfordian as is shown in Table 2.

If this extrapolation is correct and we consider the entire system of the Callovian–Oxfordian basins, the maximal temperatures about 26°C were characteristic in that system of the southern part of the Mediterranean Tethys (the present-day Egypt area). This indicated value is close to the present-day average annual water temperature near the equator. The minimal APA values (19.6°C) comparable with the temperature of warm subtropical waters would be then in the eastern Greenland area. According to accepted classification, the

Table 2. Extrapolated absolute paleotemperature analogues for the late Oxfordian

RTC	1.86	2.04	2.22	2.4	2.58	2.76	2.94	3.12	3.3	3.48	3.66	3.84
APA (°C)	19.2	19.8	20.5	21.2	21.9	22.5	23.2	23.9	24.5	25.2	25.9	26.5

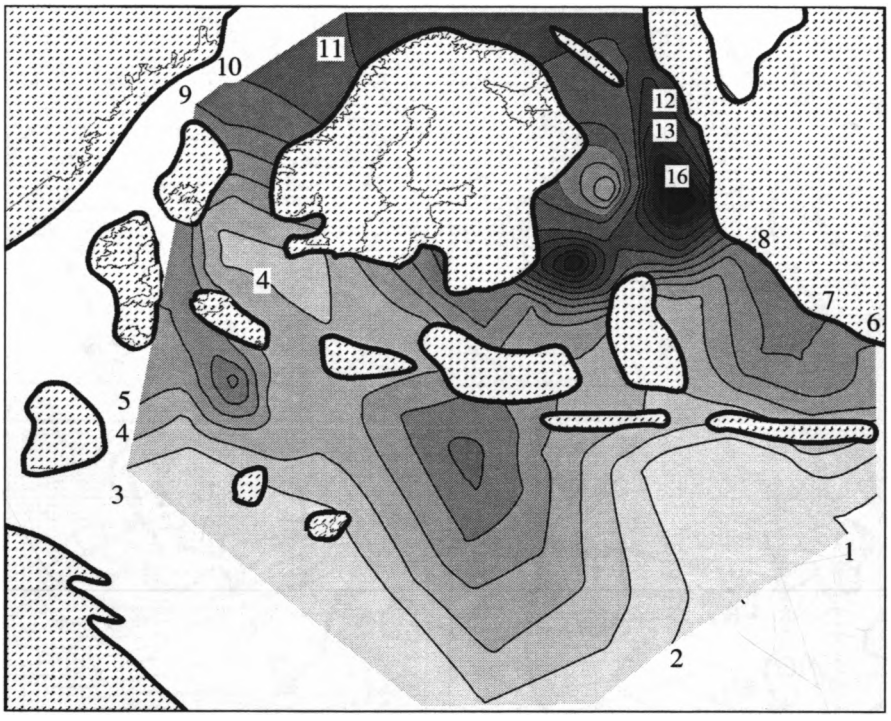

Fig. 9. Amplitudes of RPC variations values in northwestern Eurasia during the Callovian–Oxfordian period shown as mean square deviations in percents (numerals attached to contours) from the average RPC value.

boundary between tropical and subtropical water masses corresponds to the isotherm 20°C (Zhizhchenko, 1974). In our case, this isotherm corresponds to the RPI value of 2.04. The variable southernmost position of this isotherm during the early and late Callovian and in the late Oxfordian is shown in presented charts (Figs. 2–7). As one can see in figures, the greater part of northwestern Eurasia was under influence of tropical water masses during the above periods. The southernmost areas of the Boreal basin (the eastern Greenland and Pechora sea) were beyond that influence. The internal part of the Central Russian sea appeared to be northward the isotherm in question only in the late Oxfordian (Fig. 11). These inferences are consistent with absolute paleotemperatures estimated for the early–middle Callovian of England (based on ammonites only, Anderson *et al.*, 1994), being in disagreement with the data on the late Callovian–early Oxfordian basins of northwestern Europe, Greenland, European (Riboelleau *et al.*, 1998) and Central Russia (Barskov and Kiyashko, 2000). In two last works, the isotopic paleotemperature values are substantially lower, equal to 8–17.8°C (first work, fig. 2) or to 6–12.5°C (last work, fig. 1). It is difficult to comment this discrepancy because of the well-known ambiguity of isotopic paleothermometry methods. For example, Gruszczynski (1998) repudiates the paleotemperature significance for changes in geochemical parameters, if there is a positive correlation between variations in $\delta^{18}O$ and $\delta^{13}C$ values. He interprets data of radiocarbon analysis as indicative only of water stratification degree (and anoxic/oxygenated state of bottom sediments) in paleobasins. In most publications, the mentioned inclusive, the positive correlation between secular variations of $\delta^{18}O$ and $\delta^{13}C$ parameters is actually observable. In work by Anderson *et al.* (1994), a significant scatter of isotopic paleotemperature values is noted only for taxonomic groups dwelling at different depth levels. The highest paleotemperatures are characteristic of ammonites that populated the upper water layer of sea basins (15–30°C with the normal distribution optimum close to 20–23°C). For Callovian belemnites, this value is 11–19°C with the optimum approaching 13–17°C. The discrepancy is explained in the work by a much deeper habitat zone of belemnites. This factor is also a good explanation to the discrepancy mentioned higher, because paleotemperature determinations in most works are based on belemnites.

Two kinds of paleotemperature parameters, the absolute and relative, reflect approximately the same situation (types of water masses, their boundaries and distribution) for the entire system of sea paleobasins in northwestern Eurasia. This suggests that they reflect some real trends. The outlined situation can be argued for by additional data on distribution of various temperature indicators, e.g., of hermatypic corals. In the late Oxfordian, scleractinian corals migrated far northward up to the Pomor'e region and Scotland (Fig. 10). In the last area, they formed massive, although small bioherms and biostromes (Krasnov, 1983). Plotting the localities of coral buildups on the RPI chart, one can see

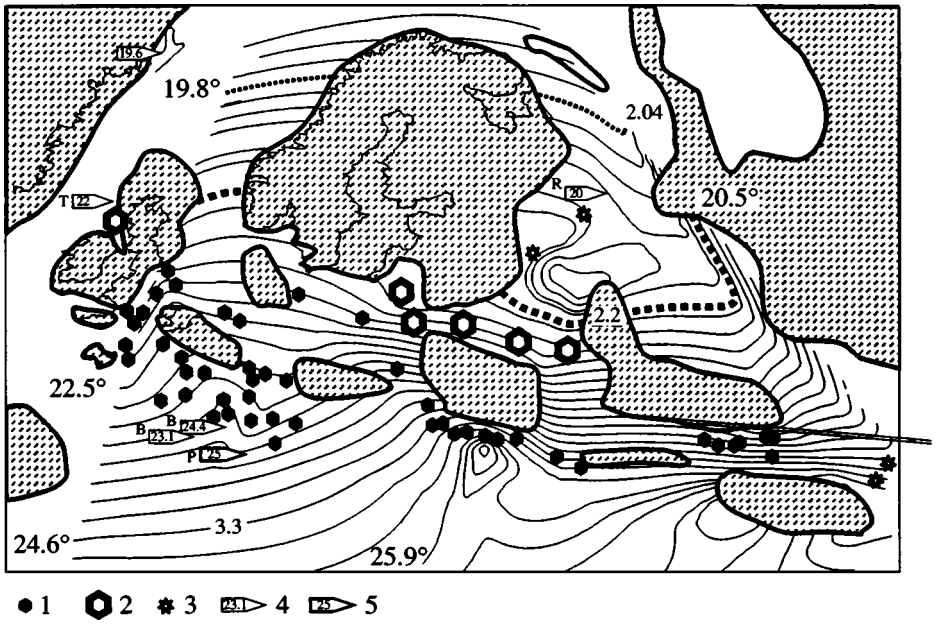

Fig. 10. The northern distribution limit of late Oxfordian scleractinian coral in northwestern Eurasia: (1) massive reefal buildings; (2) bioherms and biostromes; (3) solitary scleractinian corals or small coral meadows (after Arkell, 1933, 1961; Gerasimov, 1955; Sazonova and Sazonov, 1967; Krasnov, 1983; Leinfelder, 1993; Bertling, 1993; Bertling and Insalaco, 1998); (4) platform or periplatform and (5) Alpine thrust sites, where paleotemperatures have been estimated based on $\delta^{18}O$ in belemnites (in the last case, sampled sediments accumulated southward, and estimated values characterize southerly areas). The RPI contours are coordinated with absolute paleotemperatures in centigrade as explained in the text. Thick dotted isoline 2.2 approximately corresponds to thermal boundary (isotherm 20.5°C) that restrains the present-day distribution of hermatypic corals (see text). Thin dotted isoline 2.04 approximately corresponds to demarcation line between tropical and subtropical water masses (see text). Literal symbols designating publication and source areas of fossils: (B) Bowen, 1969 (Greenland, Miln Land; France, Deux; northern Switzerland, Aargo); (P) Padden *et al.*, in press (southern Switzerland, Montsalvens); (R) Riboulleau *et al.*, 1998 (European Russia, Makar'ev); (T) Tan *et al.*, 1970 (Scotland, Sky).

a remarkable fact: all the northernmost finds of hermatypic scleractinian corals are located near the relative paleotemperature isoline 2.22, which corresponds to the isotherm 20.5°C, the temperature boundary of the present-day distribution of reef-building corals. Moreover, almost all the buildups are located south of this isotherm, being distributed conformably to the entire isoline system. This is a distinct indication of the thermal control over distribution of the late Oxfordian corals. The observed boundary of hermatypic Scleractinia species seems to be real, because northward of the boundary, Subboreal coral assemblages are impoverished in taxonomy and ecologic aspects. In the northernmost localities (the RPI 2.7 approximately corresponding to the isotherm 22°C) of Scotland, the Pripyat trough, and Donets basin, corals form small bioherms, not reefal buildups, that is typical of suppressed corral communities. North of the isoline 2.22 (Moscow region, Unzha locality), there are known sporadic finds of corals, mainly of minute solitary but not hermatypic forms (Gerasimov, 1955). Suppressed assemblages of non-hermatypic corals are discovered also in Turkmenistan (Krasnov, 1983; Amanniayazov, 1971) at the latitude of mass occurrence of reef-building corals. It can be assumed that this regional advance southward of the suppressed coral assemblages is caused by influence of a cold current. It is clear that considered data on distribution of corals cannot be taken for the undisputable argument in favor of the suggested model, because radiation of hermatypic scleractinian corals is also controlled by other factors in addition to the temperature (Leinfelder, 1993). The causes responsible for expansion of corals in the late Oxfordian are complex, requiring special analysis. Nevertheless, the outlined zoning in distribution of hermatypic corals suggests a significant temperature influence on this biotic group.

Secular RPC changes. During the Callovian and Oxfordian ages, secular changes of RPC values are quite regular, frequently rhythmical in most areas of 54 considered above (Figs. 11, 12). In the middle Callovian, the RPC values became elevated nearly in all the areas and attained their peak level in some of them. In Subboreal (European basin and Central Russian sea) and Arctic regions, a trend of gradual or stepwise RPC decrease from the middle Callovian to the late Oxfordian is established, and the last substage is marked by a sharp fall in this parameter. The relative paleotemperature minimum at the Callovian–Oxfordian boundary time is also distinguishable in many areas. In addition to global regularities, there were characteristic regional fluctuations. For example, secular RPC variations in the Arctic and Tethyan basins were different in principle,

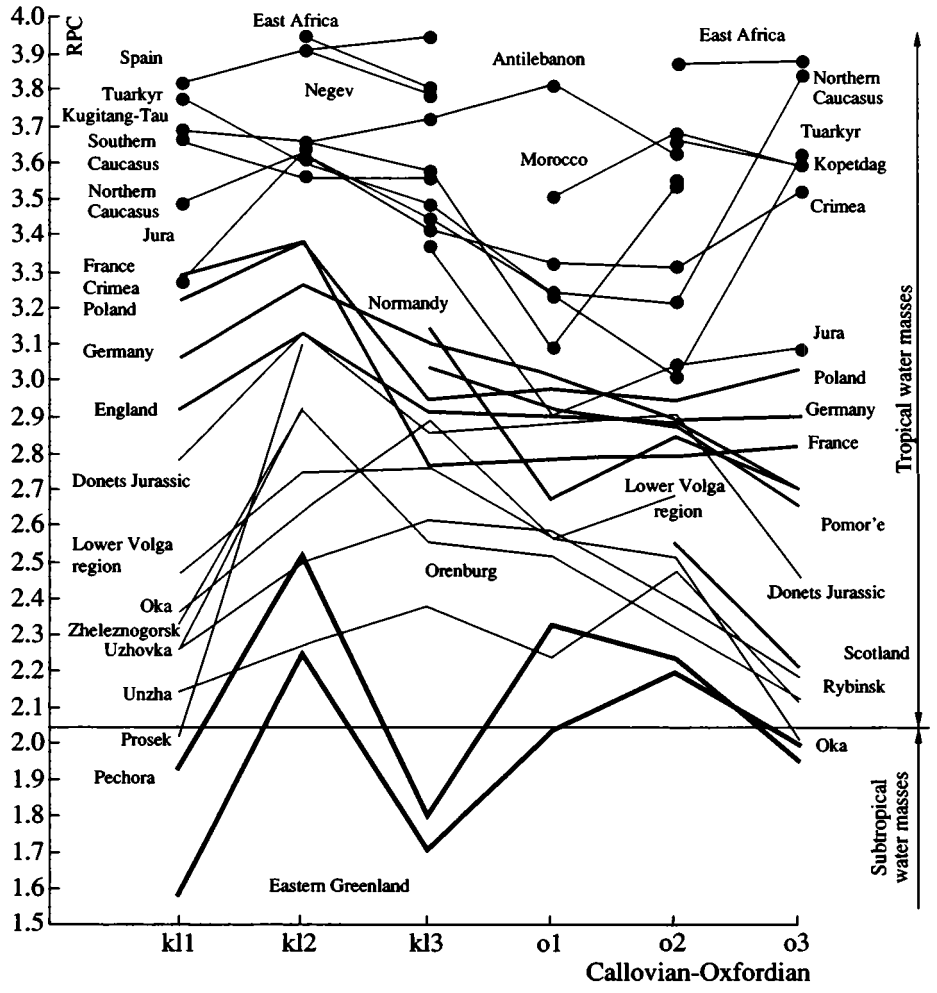

Fig. 11. Secular RPC variations in 30 areas of northwestern Eurasia during the Callovian–Oxfordian (after Kiselev, 1998, Fig. 2, with additions and modifications). Thick, medium-thick and thin lines characterize variations in the Arctic basin, European areas and Central Russian sea, respectively; lines with circles characterize Tethyan areas.

particularly during the Oxfordian. Similarly, the late Oxfordian warming in the Crimean–Caucasian sector of the Tethys (the northward advance of Tethyan current) was concurrent to cooling in the Central Russian sea (intensification and bifurcation of Boreal currents). More features in common are characteristic of Subboreal regions, although numerous deviations are observable there as well. The highest similarity between secular variation curves is characteristic of *areas of the same region.* This fact implies that regional fluctuations of thermal regimes in the Callovian and Oxfordian basins prevailed over global changes. Warming or cooling events in common likely occurred as well, though periodically. The approximated variation trends exhibit three peaks in common: the early Callovian minimum, middle Callovian maximum, and Callovian–Oxfordian boundary minimum (Fig. 12). The correlation coefficients for these events are sufficiently high only for the Arctic basin, and global significance of corresponding changes in thermal regimes can be considered only as conventional. Nevertheless, the mentioned and some other peaks detectable in general and regional variation curves are consistent with published data on oxygen isotope ratios in carbonates from several Callovian–Oxfordian sections of northwestern Europe and America. The thermal minimum of the Callovian–Oxfordian boundary time is recorded in England, Poland (Gruszczynski, 1998), and Central Russia (Riboulleau *et al.*, 1998; Barskov and Kiyashko, 2000). The late Oxfordian minimum and middle Oxfordian maximum (partly) are detected in the Peri-Tethys, western Cuba, and Central Russia (Riboulleau *et al.*, 1998; Barskov, private communication). The coincident rhythms of $\delta^{18}O$ and RPC variations seem to be not incidental and may be related to real paleotemperature fluctuations.

DISCUSSION

The considered data on circulation systems in the Callovian–Oxfordian seas of northwestern Eurasia sug-

gest that main detectable currents had stable position and did not change in quantity. These were the European and Central Russian cold effluent currents and the warm Tethyan and Dnieper currents. These currents were of approximately the same geographic position and direction.

Characteristic of the Callovian–Oxfordian time was the repeated formation of convergence zones of warm and cold currents with presumably vertical circulation and stable location. In the Central Russian sea, the convergence zone was always located in its internal part, at the junction site of three sea arms (middle Callovian to late Oxfordian). In the Tethys, the convergence zone was near the southwestern margin of the Ukrainian land, in the present-day Dniester River area (early Callovian, middle Callovian, late Oxfordian). The steady position of these elements in the circulation system implies that they characterize real events. Therefore, some deviations from the general trends should also be considered as real. The validity of suggested general model is also evident from unidirectional changes in the circulation system of Callovian–Oxfordian seas. For example, the Boreal European and warm Tethyan currents steadily migrated westward during the period under consideration. The outlined regional differences in the hydrological regime are consistent with available data on absolute paleotemperatures, which are discussed above.

The suggested circulation system is rather intricate, frequently inconsistent with factors that control the hydrological regime. For example, directions of some currents changed contrary to the Coriolis forces. To explain these deviations, I should remind that the region under consideration represented mainly a system of small epicontinental seas during the Jurassic. The direction of currents was controlled in this case by the gradient between low and high water levels in the basin. That gradient depends largely on disposition of drainage and discharge areas, on the latitude and configuration of the basin, its bottom morphology, and on location of river mouths, especially in small basins (Bowden, 1988). The Coriolis force is a factor of global circulation and affects the oceanic currents that is applicable to the Late Jurassic circulation system as well: the warm probably equatorial Tethyan current of the latter was west-oriented during the entire Callovian–Oxfordian, like in the present-day oceans. In the western terminal part of the Tethys, it met the land and deviated under the Coriolis force northward to reach Greenland, where it turned naturally eastward and probably reached Siberia (Teis and Naidin, 1973; Khudolei, 1984, 1985, 1988, 1993).

It is remarkable that practically all paleotemperature changes in the considered system of basins were of the regional, but not global character throughout the Callovian–Oxfordian time span. The relative paleotemperature variation curves plotted for separate small areas (Fig. 12) are cophasal only within a single region. The

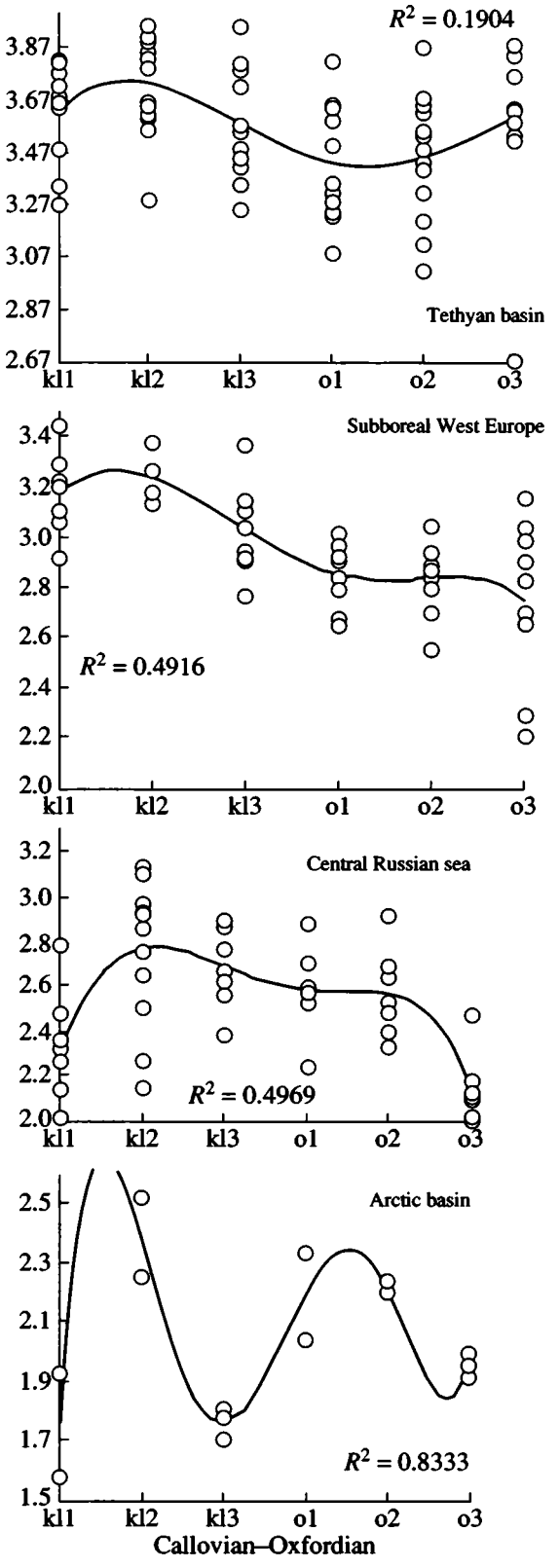

Fig. 12. Generalized trends (approximation lines correspond to polynomials risen to sixth power) characterizing secular RPC variations four Callovian–Oxfordian sea basins of northwestern Eurasia (approximated values for each region are presented in Table 1 and plotted in Fig. 11).

curves probably reflect changes in directions of warm and cold currents, which differed in various basins. Even the middle Callovian general warming of water masses was likely of a regional character, and climatic interpretation of the data obtained is hardly possible. The highest amplitudes of RPC value variations (Fig. 9) are established in areas with permanent fairways of presumable currents, and such a situation suggests that paleotemperatures primarily depend on the hydrological factor. Other factors that are able to influence directly the water temperature (solar constant, air temperature, and latitude) are almost imperceptible in the considered data. There are grounds to believe that in fact climatic changes followed the changes in the thermal regime of sea basins in question. The mutual character of thermal changes in the land climate and marine medium in the Jurassic is obvious for some regions, e.g., for the North Sea (Abbink *et al.*, 2001). Paleoclimatic curves based on palynological data from the former Scandinavian land (ibid) and entire northwestern Europe (Poulsen, *et al.*, in press) remarkably coincide with the relative paleotemperature variation curves based on ammonites from Subboreal and Boreal areas (Figs. 11, 12).

A similar dynamics of Callovian–Oxfordian transgressions and regressions in the considered system of sea basins is a probable explanation of more or less significant warming and cooling episodes. Indeed, the transgression peak that occurred in the middle Callovian (Sazonova and Sazonov, 1967; Hallam, 1988, 2001; Westermann, 1993) is correlative with the universal "warming" inferable from all paleotemperature indicators. The same situation is characteristic of the middle Oxfordian (Abbink *et al.*, 2001; Gruszczynski, 1998) that was the warmest age of the entire Jurassic period in opinion of Gruszczynski. The regression peaks of the Bathonian–Callovian and Oxfordian–Kimmeridgian boundary times were associated with decreasing paleotemperature parameters (Hallam, 1988, 2001; Norris and Hallam, 1995) that was untypical however of the regression recorded in the Callovian–Oxfordian boundary interval (Westermann, 1993; Sazonova and Sazonov, 1967). This situation can logically be explained by eustatic sea-level changes (Haq *et al.*, 1998), which are of a critical amplitude at the culmination time of transgressions and regressions.

The paleotemperature–paleohydrological situation under consideration slightly resembles the present-day thermal type of oceanic water exchange between high and low latitudes, the triggering mechanism of which is governed by sinking heavy cold waters in polar areas that form the psychrosphere. This circulation model was already referred to by consideration of the late Jurassic hydrological regime in the Viking Straits (Oshmann, 1987). In the alternative model proposed for the Jurassic and Cretaceous periods, there was suggested the other haline mechanism of water exchange (Krassilov, 1985; Miller 1991; Nesov, 1997 and others; Abbink *et al.*, 2001), when the triggering sinking of warm hypersaline waters took place in low latitudes (S-downwelling after Nesov). Despite the prevalence of epicontinental seas in northwestern Eurasia of the Callovian–Oxfordian period, the region was lacking large evaporation basins necessary for launching the haline water circulation. It cannot be ruled out however that circulation in the basins under consideration was of the intermediate type, combining elements of both the haline and thermal water exchange.

Reviewers M. A. Zharkov and N. M. Chumakov

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Genus *Berriasella* (Ammonoidea) and Ammonoid Zonation in the Berriasian of the Crimean Mountains

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Abstract—Six *Berriasella* species (Ammonoidea) found in the Berriasian sediments of the Crimean Mountains are *B. jacobi* Maz., *B. callisto* (d'Orb.), *B. subcallisto* (Touc.), *B. oppeli* (Kil.), *B. moesica* Nik. et Man., and *B. chomeracensis* (Touc.), the last two being first encountered. Based on the recognized standard zones *jacobi–occitanica–boissieri*, the Berriasian of the Crimea is correlated with coeval sections in other Mediterranean areas. Systematic description is presented for all the species.

Key words: Crimean Mountains, Berriasian, ammonites, zonal biostratigraphy, correlation.

INTRODUCTION

Being of a very wide geographic range, the extremely variable *Berriasella* species are most important for zonal subdivision of the Tithonian–Berriasian transition and for determination of the Jurassic–Cretaceous boundary in the Mediterranean region. The species are described in many publications (Mazenot, 1939; Le Hégarat, 1973; Nikolov, 1982; Tavera, 1985; etc.) devoted to Tithonian and Berriasian stratigraphy of southern Europe (France, Spain, and Bulgaria). Based on these investigations, the *Berriasella jacobi* Zone of the corresponding index species is now established to be the lower Berriasian zone in Mediterranean areas.

BERRIASELLA SPECIES AND THEIR SIGNIFICANCE FOR THE BERRIASIAN AMMONOID ZONATION

Mazenot (1939) who understood genus *Berriasella* in a very broad sense described among other the species *B. jacobi* Maz., *B. oppeli* (Kil.), *B. callisto* (d'Orb.), *B. subcallisto* (Touc.), *B. oxycostata* (Jacob), *B. picteti* (Jacob), *B. privasensis* (Pict.), and *B. chomeracensis* (Touc.), which belong to the genus in its modern interpretation.

Le Hégarat (1973) who studied the stratotype section in southeastern France described 17 *Berriasella* species. He classed them with subgenera *Berriasella* and *Picteticeras*. The corresponding species listed in his work are *(B.) callisto* (d'Orb.), *B. jacobi* Maz., *B. moreti* Maz., *B. oppeli* (Kil.), *B. paramacilenta* (Maz.), *B. privasensis* (Pict.), *B. sabatasi* Le Hégarat), *P. aurousei* Le Hégarat, *P. chomeracensis* (Touc.), *P. elmii* Le Hégarat, *P. enayi* Le Hégarat, *P. evoluta* (Le Hégarat), *P. jauberti* (Maz.), *P. moesica* (Nik. et Man.), *P. oxycostata* (Jacob), and *P. picteti* (Jacob). Le Hégarat reported that *B. jacobi* Maz., *B. moreti* Maz., *B. oppeli* (Kil.), *B. sabatasi* Le Hégarat, *P. aurousei* Le Hégarat, *P. chomeracensis* (Touc.), *P. elmii* Le Hégarat, and *P. enayi* Le Hégarat, are mainly characteristic of the *jacobi* Zone of his zonation chart (table), whereas the others are typical of the younger *"grandis"–occitanica–boissieri* zones of the stratotype. *B. callisto* (d'Orb.) is stratigraphically the youngest species among the listed forms. It was described by Le Hégarat from the *picteti* and *callisto* subzones of the *boissieri* Zone.

Tavera (1985) studied in detail the ammonite distribution througout the Tithonian and Berriasian sections of southern Spain and described 16 *Berriasella* species. Among them, 11 species *B. jacobi* Maz., *B. oxycostata* (Jacob), *B. elmii* Le Hégarat, *B. mazenoti* Breist., *B. aurousei* Le Hégarat, *B. chomeracensis* (Touc.), *B. subcallisto* (Touc.), *B. berthei* (Touc.), *B. moreti* Maz., *B. sabatasi* Le Hégarat, and *B. oppeli* (Kil.) are mainly characteristic of the *jacobi* Zone, as it was defined by Tavera (table). *B. privasensis* (Pict.) and *B. (Hegaratella) paramacilenta* (Maz.) were attributed by Tavera to the overlying *andrussowi* Zone, while *B. callisto* (d'Orb.) and *B. picteti* (Jacob) were regarded as characteristic of the *boissieri* Zone. From the Tithonian *Durangites* Zone, Tavera described the new species *B. tithonica* Tavera.

Numerous *Berriasella* forms are known from the Tithonian–Berriasian boundary beds in Bulgaria (Nikolov, 1982). Nikolov stated that the *Transitorius* Zone of the upper Tithonian is the lowermost unit with *Berriasella* species. It is remarkable however that describing a 0.5-m-thick layer of the section he reported that it yields *Paraulacosphinctes transitorius* (Opp.), *Berriasella (picteticeras) chomeracensis*

Zonal subdivision of the Tithonian–Berriasian boundary sediments in the Crimean Mountains and their interregional correlation

Southeastern France, Le Hégarat, 1973			Southeastern France, *Colloque...*, 1975			Bulgaria, Nikolov, 1982				Spain, Tavera, 1985			Tunisia, Tavera *et al.*, 1986			Crimean Mountains, after the authors			
System	Stage	Zone	System	Stage	Zone	System	Stage	Zone		System	Stage	Zone	System	Stage	Zone	System	Stage	Zone	
Cretaceous	Berriasian	*Occitanica*	Cretaceous	Berriasian	*Occitanica*	Cretaceous	Berriasian	*Occitanica*		Cretaceous	Berriasian	*Andrussowi*	Cretaceous	Berriasian	*Neocosmoceras*	Cretaceous	Berriasian	*Occitanica*	
		Grandis			*Jacobi-grandis*			*Grandis*				*Jacobi*			*Occitanica*			*Jacobi*	*Grandis*
Jurassic	Tithonian	*Jacobi*	Jurassic	Tithonian	?	Jurassic	Tithonian	*Transitorius*	*Chaperi*						Grandis				
															Pseudogroteanum				*Chomeracensis*
															Chomeracensis				
					Transitorius				*Microcanthus*	Jurassic	Tithonian	*Durangites*	Jurassic	Tithonian	?	Jurassic	Tithonian	?	
												Transitorius							

(Touc.), *Malbosiceras chaperi* (Pict.), and *Pseudosubplanites euxinus* (Ret.) (Nikolov, 1982, p. 16). This assemblage includes most likely the redeposited ammonites of different age, since neither in Spain nor in France *P. transitorius* (Opp.) was found in association with *B. chomeracensis* (Touc.) and *P. euxinus* (Ret.). Accordingly, as we think, the *chaperi* Subzone was erroneously placed into the *Transitorius* Zone. Sediments with *Malbosiceras chaperi* and *Delphinella delphinensis* should be younger but not coeval to the *P. transitorius* beds. This was proved by Egoyan (1975). In our opinion, the *chaperi* Subzone of the *Transitorius* Zone distinguished by Nikolov in Bulgaria is actually a part of the Berriasian *jacobi* Zone established in France and Spain (Hoedemaeker and Bulot, 1990), i.e. it should be joined in Bulgaria with the *grandis* Zone. Mixing of ammonites of different age in a single thin bed can be caused by various reasons. Therefore, the *Berriasella* species first appear in Bulgaria not earlier than in the Berriasian.

According to many authors (Behrendsen, 1891; Gerth, 1925, 1926; Krantz, 1926; Leanza, 1945), the *Berriasella* species occur also in Argentina. However, we figured out that most of the described taxa are not the *Berriasella* species proper, as they have either tripled ribs or prominent node-like inflations at the rib branching points or at the ventrolateral shoulders. The true *Berriasella* species, in our opinion, are *B. oppeli* (Kilian) described by Behrendsen (1891) and *B. krantzi* Leanza (Leanza, 1945). Behrendsen probably figured a macroconch *B. oppeli* having D = 89 mm, whereas specimens of this taxon from Western Europe are usually represented by microconchs. In general, *Berriasella* froms from Argentine show endemic traits.

In the upper Tithonian *Substeueroceras* beds of Mexico, there was found *B. zacatecana* Imlay (Imlay, 1939). *B.* cf. *picteti* (Kilian) and *B.* aff. *zacatecana* Imlay are known from the Berriasian of Mexico (Verma and Westermann, 1973), however, their specimens are poorly preserved, being unsuitable for a reliable correlation with the West European forms.

A diverse *Berriasella* assemblage dominated by West European species is known from northern Africa (Tunisia) and Madagascar. Species described in Tunisia (Arnould-Saget, 1951) are *B. oppeli* (Kilian), *B. privasensis* (Pictet), *B. oxycostata* (Jacob), *B. subcallisto* (Toucas), *B. jacobi* Mazenot, *B. moreti* Mazenot, *B. chomeracensis* (Toucas), *B. pseudomoravica* Arnould-Saget, and *B. constricta* Arnould-Saget. Their assemblage is characteristic of the Berriasian *jacobi–occitanica* Zone.

The Berriasian sediments of Madagascar yield *B. privasensis* (Pictet), *B. jacobi* Mazenot, *B. callisto* (d'Orbigny), *B. picteti* (Jacob), and *B. besairiei* Collignon (Collignon, 1960,1962). Accordingly, the Berriasian *jacobi, occitanica*, and *boissieri* zones seem to be distinguishable there.

Species *B. (B.)* cf. *oxycostata* (Jacob), *B. (B.)* aff. *chomeracensis* (Toucas), *B. (B.) oppeli* (Kilian), and *B. (Elenaella) sevenieri* (Le Hégarat) are known from the Berriasian of Yemen (Howarth and Morris, 1998). *B.* aff. *chomeracensis* cannot be referred to the genus *Berriasella* because of tripled ribs, and *sevenieri* form in modern taxonomic nomenclature belongs to the genus *Delphinella. B. (B.) oppeli* is represented by two fragments, which are inappropriate, in our opinion, even for identification at the generic level. The *jacobi* Zone is

evidently present in sections of Iraq, since *B. (B.) jacobi* Mazenot was identified there (Howarth, 1992).

In the Himalayas, where Uhlig (1903–1910) found *Hoplites (Berriasella)* cf. *privasensis* (Pictet), a diverse *Berriasella* assemblage was recently encountered (Liu and Wang, 1987). Chinese researchers recognized here with confidence the *jacobi* Zone that yields species *B. jacobi* Mazenot, *B.* cf. *sabatasi* Le Hégarat, *B. oppeli* (Kilian), *B. chomeracensis* (Toucas), *B. elmii* Le Hégarat, *B. aurousei* Le Hégarat, *B. subcallisto* (Toucas), *B. besairiei* Collignon, and *B. xizangensis* Liu. Judging from illustrations, the Himalayan *Berriasella* species are rather close to the West European forms.

Among ammonites from New Guinea (Helmcke *et al.*, 1978) there was found *B. (picteticeras) moesica* (Nik. et Man.), but its figure has not been published.

Sey and Kalacheva (1999) who studied sections in the southern Primor'e, the Far East of Russia, described ammonites of genera *Dalmasiceras, Pseudosubplanites*, and *Berriasella* from sediments bearing numerous *Buchia* forms of the upper Volgian *Buchia piochii* s.l.–*B. terebratuloides* Zone. In their opinion, the studied ammonites can be attributed the *jacobi* Zone of the Tethyan Berriasian. It is undoubtedly an interesting opinion, however, we should note that all ammonites were identified in the open nomenclature (*Berriasella* ex gr. *jacobi* Maz., *Berriasella?* sp., *Pseudosubplanites* cf. *grandis* (Maz.), *P.* aff. *combesi* Le Hégarat, and *Pseudosubplanites* sp.), and *Dalmasiceras orientale* is a new species. Our revision of ammonites collected by Sey and Kalacheva (collection is deposited in the Central Research Geological Prospecting Museum, St. Petersburg) showed that shell shape and ribbing of *B.* ex gr. *jacobi* are consistent with the *Berriasella* diagnosis (microconchs with double ribs). At the same time, the shell is devoid of the ribs ventral break characteristic of the genus, and only a slight rib lowering without an interruption is visible in the mid-ventral side. Shells of *Berriasella?* sp. cannot be referred, in our opinion, to the genus *Berriasella* at all.

Berriasella species from the Berriasian of the Northern Caucasus and Transcaucasia (Georgia) are as yet imperfectly studied. Khimshiashvili (1976) reported on occurrence of *B. callisto* (d'Orb.), *B. subcallisto* (Touc.), and *B. privasensis* (Pict.) in the Transcaucasia, but only the latter species is described and illustrated in this work. Later on, *B. jacobi* Maz., *B. oxycostata* (Jacob), and *B. privasensis* (Pict.) from the Berriasian of Abkhaziya have been described and figured (Khimshiashvili, 1989). Two former species are interpreted as characteristic of the *jacobi* Zone, and the latter one is attributed to the *occitanica* Zone.

Sakharov (1984) identified and illustrated (but not described) a single species *Berriasella* cf. *callisto* (d'Orb.) from the Urukh River section of the northern Caucasus. Unfortunately, only the lateral view of shell is reproduced. The specimen is a fragment of phragmocone, and it is difficult to define, in our opinion, if it is related by affinity to the genus *Berriasella*. Sey and Kalacheva studied the Berriasian ammonites from the Urukh section (Kolpenskaya *et al.*, 2000), among which they identified *Berriasella (Picteticeras)* cf. *evoluta* Le Hégarat, *B. (P.)* cf. *jauberti* (Maz.), *B. (P.)* sp., and *B. (B.) callisto* (d'Orb.). The species were encountered along with representatives of *Euthymiceras, Spiticeras, Subalpinites*, and *Riasanites* genera. In opinion of Sey and Kalacheva (p. 26), the upper Berriasian part of the Urukh section distinguished as the *Berriasella callisto–Jabronella paquieri* Beds can be correlated with the *callisto* Subzone of the *boissieri* Zone of the stratotype.

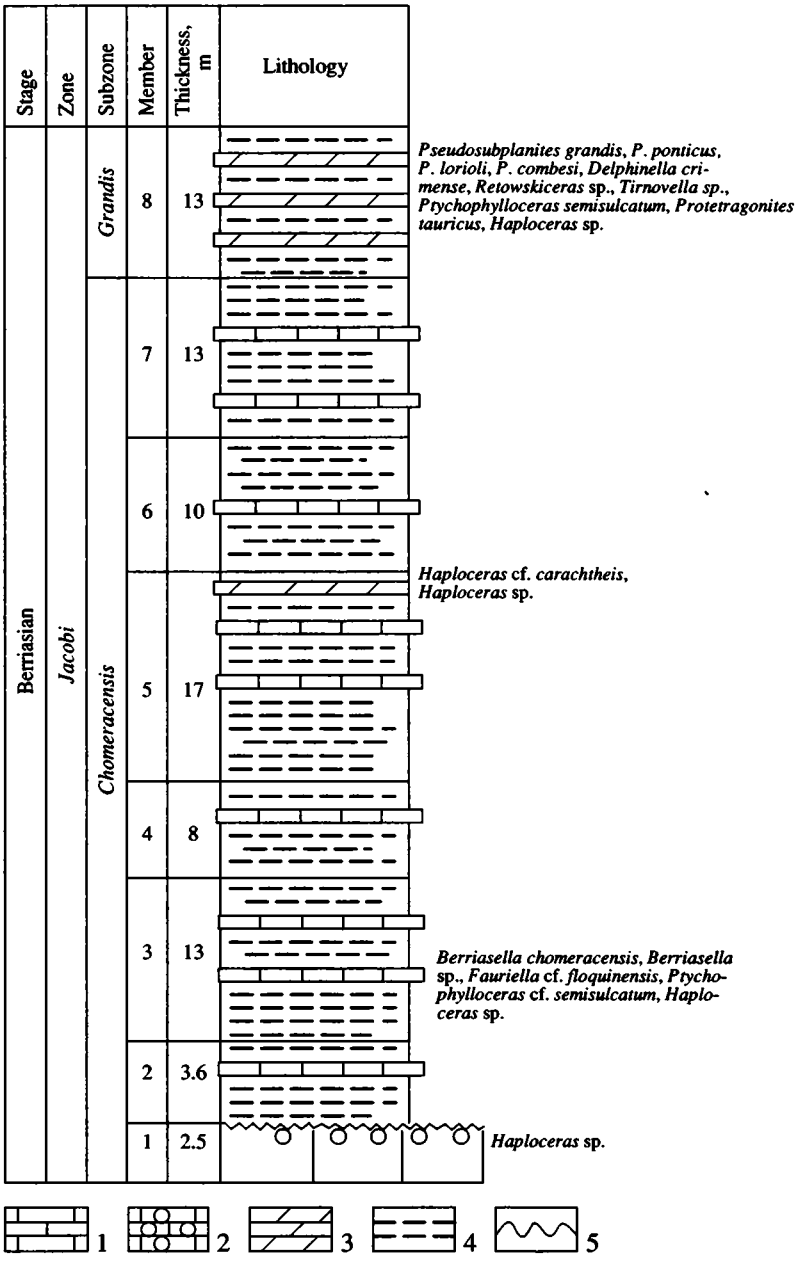
Berriasella species from the Crimea are mentioned in a series of publications (Sokolov, 1886; Retowski, 1893; Drushchits, 1960; Lysenko and Popov, 1962; Uspenskaya, 1969; Druschits, 1975; Kvantaliani and Lysenko, 1979; Lysenko and Yanin, 1979; Bogdanova *et al.*, 1981, 1984, 1999; Glushkov, 1997; Arkad'ev, 2002; Arkad'ev and Savel'eva, 2002). According to the present notion, the Crimean *Berriasella* assemblage is characteristic of the *jacobi–boissieri* Zone. However, the frequently occurring species of the assemblage are inadequately studied so far, being attentively considered in a few works only.

MATERIALS

The studied *Berriasella* specimens have been collected in various periods by Drushchits, Lysenko, Nerodenko, Yanin, Bogdanova, Lobacheva, Prozorovskii, Favorskaya, Glushkov, Savel'eva, and Arkad'ev from Tithonian–Berriasian sediments of the Crimean Mountains. When studying the collection, we identified species *B. jacobi* Maz., *B. moesica* Nik. et Mand., *B. subcallisto* (Touc.), *B. oppeli* (Kil.), *B. callisto* (d'Orb.), and *B. chomeracensis* (Touc.) described below. The collection is stored at the Central Research Geological Prospecting Museum in St. Petersburg (no. 13098).

Four *Berriasella* species (*jacobi, oppeli, subcallisto*, and *callisto*) were known previously, while the other two represent new forms for the Crimea. Most of the species are referred to the *jacobi* Zone recognizable in sections of Feodosia, eastern Crimea, and the Tonas and Sary-Su River basins. Two specimens of *Berriasella callisto* (d'Orb.) characteristic of the *boissieri* Zone (Le Hégarat, 1973) were found in the central Crimea near the Mramornoe Village, in a separate tectonic block of Berriasian sediments.

The most representative sections of the Tithonian–Berriasian boundary sediments are known in the vicinity of Feodosia. For instance, the section on the Svyatogo Il'i Cape has been studied since the end of the XIX century (Retowski, 1893; Druschits, 1975; etc.). The section located 200 m west of the cape is 80 m thick in total. The famous "Feodosia Marl" bearing the ammonite assemblage described by Retowski represents the uppermost member of the section and is 13 m



Ammonites in the Berriasian section on the Svyatogo Il'i Cape (Feodosia): (1) limestone; (2) brecciated limestone; (3) marl; (4) clay; (5) unconformity.

thick. According to works mentioned above, the marl yields the following species: *Euphylloceras serum* (Opp.), *Macrophylloceras beneckei* (Zitt.), *Ptychophylloceras inordinatum* (Touc.), *P. semisulcatum* (d'Orb.), *Holcophylloceras tauricum* (Ret.), *Biasaloceras liebigi* (Opp.), *Protetragonites tauricus* (Kulj.-Vor.), *Haploceras carachtheis* (Opp.), *H. elimatum* (Opp.), *Haploceras* sp., *Substreblites zonarius* (Opp.), *Spiticeras orientale* (Kil.), *S. proteus* (Ret.), *S.* ex gr. *theodosia* (Desh.), *Berriasella oppeli* (Kilian), *B. (?) subcallisto* (Touc.), *Berriasella (?)* sp., *Pseudosubplanites lorioli* (Zit.), *P. grandis* (Maz.), *P. combesi* Le Hégarat, *P. ponticus* (Ret.), *P. subrichteri* (Ret.), *Delphinella berthei* (Touc.), *D. delphinensis* (Kil.), *D. subchaperi* (Ret.), *D. crimense* (Burckh.), *D. obtusenodosa* (Ret.), *Fauriella* ex gr. *shipkovensis* Nikol. et Man., *Retowskiceras andrussowi* (Ret.), *Retowskiceras* sp., and *Tirnovella* sp.

Pseudosubplanites and *Delphinella* species present in the assemblage suggest that the marly member is certainly correlative with the standard Berriasian *jacobi* Zone. The distribution of ammonites throughout this zone has not been studied yet in the Crimean Mountains, as below the Feodosia Marl in the Svyatogo Il'i Cape zonal ammonite species, except for *Ptychophylloceras* and *Haploceras* inappropriate for age diagnosis, were not encountered until recently (Bogdanova *et al.*, 1984). In 2001, Arkad'ev and Savel'eva (2002) found ribbed ammonites (berriasellids) in the lowermost part of the section, about 6 to 8 m above the top of the 2-m-thick marker bed of brecciated limestone that is exposed immediately near the water edge and has been considered previously (Druschits, 1975) as the Tithonian–Berriasian boundary horizon (figure). In the 13-m-thick member of greenish gray compact to splinter clay interbedded with ferruginate brown to light brown compact detrital limestones, the following ammonites were identified: *Ptychophylloceras* cf. *semisulcatum* (d'Orb.), *Haploceras* sp., *Berriasella chomeracensis* (Toucas), *Berriasella* sp., and *Fauriella* cf. *floquinensis* Le Hégarat. Dominant form of the assemblage is *Berriasella chomeracensis* (Touc.). In the upper Feodosia Marl member, it is missing. The species was described from the *jacobi* and *grandis* zones of southeastern France (Le Hégarat, 1973) and from the *jacobi* Zone of Spain (Tavera, 1985). In Bulgaria, Nikolov (1982) established a somewhat wider range of this taxon, from the upper Tithonian *Transitorius* to the Berriasian *boissieri* Zone. However, as is noted above, it most likely corresponds to the *jacobi–boissieri* Zone, thus being confined to the Berriasian. *Fauriella floquinensis* was described from the *jacobi–grandis–occitanica* zones (Le Hégarat, 1973), however the Crimean specimens are poorly preserved and unsuitable for the correct identification.

The above ammonite assemblage from lower beds of the Svyatogo Il'i Cape section is most likely indicates the Berriasian age, correlative with that of the *jacobi* Zone. Characteristic Tithonian ammonites have not been found yet at this site. Based on distribution patterns of ammonite, Arkad'ev subdivided the *jacobi* Zone of the studied section in two, the lower *Berriasella chomeracensis* (figure, Members 1–7) and the upper *Pseudosubplanites grandis* (Figure, Member 8) subzones (Arkad'ev and Savel'eva, 2002).

SYSTEMATIC DESCRIPTION

Family Berriasellidae Spath, 1922

Genus *Berriasella* Uhlig, 1905

(=*Stenoceras* Uhlig, 1911 (non d'Orbigny, 1849); *picteticeras* Le Hégarat, 1973)

Type species: *Ammonites privasensis* Pictet, 1867; Berriasian, southeastern France.

Diagnosis. Discoid shell with slightly convex flanks and a rounded venter; umbilicus wide, stepped; ribbing thin, represented by bifurcating and simple ribs. On the venter, ribs are interrupted by a furrow, clearly visible on juvenile whorls and commonly indistinct (to complete disappearance) on the last whorl.

Species composition and distribution: *Berriasella tithonica* Tavera – upper Tithonian, *Durangites* Zone in Spain; *B. zacatecana* Imlay – upper Tithonian of Mexico; *B. privasensis* (Pictet) – Berriasian, *occitanica* Zone in southeastern France, Tunisia, Algeria, Bulgaria, Caucasus, and Himalayas, Berriasian in Madagascar; *B. callisto* (d'Orbigny) – Berriasian, *boissieri* Zone (*picteti, callisto* subzones) in the Crimea, France, Italy, and Bulgaria, Berriasian in Madagascar, Berriasian *boissieri* Zone and Valanginian, *Otopeta* Zone in Spain; *B. subcallisto* (Toucas) – Berriasian, *jacobi–occitanica* zones (*subalpina* Subzone) in the Crimea, Bulgaria, and France, *jacobi* Zone in Spain and Himalayas, Berriasian in the Caucasus, Algeria, and Tunisia; *B. chomeracensis* (Toucas) – Berriasian, *jacobi* Zone (*chomeracensis* Subzone) in the Crimea, *jacobi* Zone in southeastern France, Spain, and Himalayas, *jacobi–boissieri* zones in Bulgaria, *chomeracensis* Zone in Tunisia; *B. oppeli* (Kilian) – Berriasian, *jacobi–occitanica* zones in the Crimea and Bulgaria, *jacobi* Zone in Spain, southeastern France, and Himalayas, Berriasian in Silesia, Romania, Tunisia, and Argentina(?); *B. jacobi* Mazenot – Berriasian, *jacobi* Zone in the Crimea, Caucasus, Bulgaria, southeastern France, Spain, Iraq, and Himalayas, *chomeracensis–pseudogroteanum* zones in Tunisia, Berriasian in Madagascar; *B. moreti* Mazenot – Berriasian, *jacobi* Zone in southeastern France, Spain, Bulgaria, and Tunisia; *B. elmii* Le Hégarat – Berriasian, *jacobi* Zone in southeastern France, Bulgaria, and Himalayas; *B. aurousei* Le Hégarat – Berriasian, *boissieri* Zone in southeastern France, Bulgaria, and Himalayas; *B. sabatasi* Le Hégarat – Berriasian, *jacobi* Zone in southeastern France; *B. enayi* Le Hégarat, Berriasian, *jacobi* Zone in southeastern France and Bulgaria; *B. evoluta* Le Hégarat – Berriasian, *boissieri* Zone (*paramimounum–picteti* subzones) in southeastern France; *B. oxycostata* (Jacob) – Berriasian, *jacobi* Zone in southeastern France, Tunisia, Romania, Bulgaria, and Caucasus; *B. picteti* (Jacob) – Berriasian, *boissieri* Zone (*picteti* Subzone) in France, Bulgaria, and Madagascar, Berriasian in Mexico; *B. mazenoti* Breistroffer – Berriasian, *jacobi* Zone in southeastern France and Bulgaria, Berriasian in Tunisia; *B. vranensis* Nikolov – Berriasian, *boissieri* Zone in Bulgaria; *B. subeudichotoma* Nikolov – Berriasian, *jacobi* Zone in Bulgaria; *B. moesica* Nikolov et Mandov – Berriasian, *jacobi–boissieri* zones (*paramimounum* subzone) in Bulgaria and France, *jacobi–occitanica* zones in the Crimea, Berriasian in New Guinea (?); *B. fitchevi* Nikolov – Berriasian, *boissieri* Zone (*picteti* Subzone) in Bulgaria; *B. bebrovensis* Nikolov – Berriasian, *boissieri* Zone (*picteti* Subzone) in Bulgaria; *B. preslavensis* Nikolov – Berriasian, *occitanica* Zone in Bulgaria; *B. momtchilovi* Nikolov – Berriasian, *boissieri* Zone (*callisto* Subzone) in Bulgaria; *B.*

pseudomoravica Arnould-Saget – Berriasian in Tunisia; *B. constricta* Arnould-Saget – Berriasian in Tunisia; *B. krantzi* Leanza – Berriasian in Argentina; *B. xizangensis* Liu – Berriasian, *jacobi* Zone in Himalayas; *B. besairiei* Collignon – Berriasian in Madagascar.

Remarks. The genus *Berriasella* was initially considered in a broad sense. Mazenot (1939) included in it the forms with tubercles and with fasciculate and polygyrate ribs. After subsequent revision, these ammonites were mainly excluded from the genus and distinguished as separate genera: *Fauriella* Nikolov, 1966; *Tirnovella* Nikolov, 1966; *Strambergella* Nikolov, 1966; *Delphinella* Le Hégarat, 1973; *Pseudosubplanites* Le Hégarat, 1973; *picteticeras* Le Hégarat, 1973; and *Hegaratella* Nikolov et Sapunov, 1977. The distinguished genera are unclearly distinct or close in morphology. Because of numerous transitional forms, it is difficult to discriminate these genera. Researchers often referred the same species to different genera. That is why Tavera (1985) included some problematic taxa into the initial genus *Berriasella* as *Berriasella* subgenera, e.g., *Berriasella* s.s., *Hegaratella, Pseudosubplanites*, along with the new subgenus *Busnardoiceras*. Tavera considered the genus *Strambergella* as synonym of *Fauriella* and *Picteticeras* synonym of *Berriasella (Berriasella)*. However, this approach to taxonomy of berriasellids faild to determine clear distinctions between the genera in question.

Examination of numerous papers devoted to *Berriasella* forms leads to conclusion that researchers refer sometimes the forms with simple (double and single) and composite (polygyrate or fasciculate) ribbing to the same taxon, and this complicates classification further. In our opinion, only the species with bifurcate and solitary ribs should be included into the genus *Berriasella*.

One more important diagnostic feature of *Berriasella* species is the siphonal furrow on the venter. According to our experience and published data, the furrow on the venter is changing in morphology during the ontogeny.

Berriasella forms from Western Europe are of a medium, rarely, large size. A furrow on the venter is usually distinct on juvenile whorls, being either retained or indistinct to complete disappearance on adult whorls. The Crimean *Berriasella* species we studied are small to medium in size. Among them, there are specimens with distinct furrow (microconchs) and those showing its gradual disappearance from juvenile toward last whorls. According to our observations, a furrow on the venter can be seen (1) as a narrow deep groove breaking the ribs, (2) as a shallow flute touching the upper part of ribs without a break, and (3) as a weak lowering of ribs in the mid-venter without their interruption. There are gradual transitions between these distinctive features observable sometimes in the same specimen. As the available Crimean specimens are poorly preserved (deformed internal moulds lacking shelly layer), we find it difficult to assess ultimately the taxonomic significance of the ventral furrow. Thus all the attempts of using this feature to divide the genus *Berriasella* in two groups of shells with or without furrow (Mazenot, 1939; Druschits, 1960) were unsuccessful. In our opinion, the ventral furrow is mainly characteristic of inner whorls of *Berriasella* species and, to a lesser extent, of outer ones.

In general, we agree with diagnosis of the genus *Berriasella* described by Wright *et al.* (1996) in their "Treatise on Invertebrate Paleontology," where bifurcate and single ribs and a ventral furrow on inner whorls are regarded as diagnostic features.

Considering the genus *Berriasella*, Tavera (1985) distinguished subgenus *Busnardoiceras* with the type species *Parapallasiceras busnardoi* Le Hégarat, which, unlike the true *Berriasella* species, shows polygyrate and fasciculate ribbing. Other researchers described similar forms as different *Berriasella* species. In our Crimean collection, there are such forms as well. From our standpoint, they and similar forms should be attributed to the genus *Busnardoiceras* Tavera, as the above distinctive features clearly differ it from other generic taxa.

Berriasella jacobi Mazenot, 1939

Plate, figs. 1 and 2

Hoplites carpathicus Toucas, 1890, p. 602, Pl. 17, figs. 10 and 11.

Hoplites callisto Toucas, 1890, p. 600, Pl. 17, fig. 3.

Berriasella jacobi Mazenot, 1939, p. 54, Pl. 4, figs. 1–5; Arnould-Saget, 1953, p. 33, Pl. 3, fig. 11 (not Pl. 4, fig. 1); Collignon, 1960, Pl. 165, figs. 668 and 669; Le Hégarat et Remane, 1968, p. 25, Pl. 5, figs. 1 and 2; Liu and Wang, 1987, Pl. 1, fig. 6, Pl. 2, fig. 7; Khimshiashvili, 1989, p. 9, Pl. 4, figs. 3 and 4.

Berriasella (Berriasella) jacobi Le Hégarat, 1973, p. 56, Pl. 6, figs. 9–12, Pl. 38, figs. 3, 6, and 7; Sapunov, 1979, p. 173, Pl. 55, fig. 3; Nikolov, 1982, p. 51, Pl. 8, figs. 4–8; Tavera, 1985, p. 238, Pl. 33, figs. 7–9, 11 (not fig. 10).

Shape. Shell small, semi-involute, moderately inflated, with gently convex flanks; venter wide, slightly rounded; whorl section rounded-rectangular; umbilicus wide, stepped, with steep wall.

Sculpture. Flanks are ornamented with bifurcate and rarely simple ribs. Among about 40 inner ribs present on the last whorl, there are three to four simple ribs. Originating on the umbilical wall, ribs are posteriorly deflected a little on umbilical shoulder and cross the flank almost straightly or slightly incurved anteriorly. Towards the aperture, ribs are incurved stronger. On mid-whorl or slightly above ribs bifurcate forming equal branches, with the posterior one slightly deflected backward.

In various specimens, the ventral furrow is of different morphology. In specimen 1/13098, an indistinct furrow runs across the ribs but does not interrupt them

at the beginning of the last whorl, while at the end of the whorl there is a distinct narrow furrow interrupting the ribbing. In contrast, juvenile whorls of specimens 2/13098 and 3/13098 have a distinct furrow that becomes obscure closer to aperture, where ribs cross the venter at a right angle.

Dimensions (mm) and number of ribs per whorl

Specimen	D	H	D_u	W	H/D	D_u/D	W/D	Rib number	
								outer	inner
2/13098	28.5	13.5	–	10	0.47	–	0.35	–	–
1/13098	30	12.5	8.5	9.5	0.42	0.28	0.32	–	40
3/13098	32	13	10	–	0.41	0.31	–	78	41
4/13098	36.5	13.5	12.5	–	0.37	0.34	–	–	–

Comparison and remarks. The described specimens, especially no. 1/13098, are most similar to forms figured by Mazenot (1939) and Sapunov (1979). *B. jacobi* Maz. differs from other *Berriasella* species by a relatively more inflated shell. In one of the specimens figured by Tavera (1985, Pl. 33, fig. 10), a bidichotomous rib is clearly visible on the adult whorl. Ribbing of this kind is atypical of the genus *Berriasella*, and we did not include that specimen in synonymy.

Distribution. Berriasian, the *jacobi* Zone in the Crimea, Caucasus, Bulgaria, southeastern France, Spain, Iraq, and Himalayas; *chomeracensis–pseudogroteanum* zones in Tunisia; Berriasian in Madagascar.

Material. Five specimens (1–5/13098) from the central (Sary-Su River) and eastern (Feodosia region, Barakol'skaya Valley) Crimea, collected by Druschits and Yanin.

Berriasella moesica Nikolov et Mandov, 1967
Plate, figs. 6 and 7.

Berriasella moesica: Nikolov et Mandov, 1967, p. 41, Pl. 1, figs. 1 and 3 (not figs. 2 and 4).

Berriasella (Berriasella) moesica: Le Hégarat, 1973, p. 77, Pl. 8, fig. 8; Pl. 39, figs. 10 and 11.

Berriasella (Picteticeras) moesica: Nikolov, 1982, p. 71, Pl. 16, figs. 1–6.

Shape. Discoid shell with wide, almost parallel flanks and narrow-rounded venter; cross-section of the last whorl strongly compressed, highly extended; umbilicus wide, stepped, with steep wall.

Sculpture. Flanks are ornamented with thin, dense, mostly bifurcate ribs (58–60 inner ribs per a half whorl). Ribs single or intermediate (starting from mid-whorl) are rare (1–2 per whorl). Originating on umbilical wall, ribs straightly cross flanks, being slightly incurved in the middle closer to aperture. At about upper one third of flank, ribs bifurcate into equal branches. Aperture is preserved in specimen 6/13098. A narrow furrow interrupting ribs is on the venter.

Dimensions (mm) and number of ribs per half a whorl

Specimen	D	H	W	D_u	H/D	W/D	D_u/D	Number of ribs	
								outer	inner
6/13098	42.5	18	–	–	0.42	–	–	58	30

Comparison and remarks. In distinction from other *Berriasella* species, ribs of *B. moesica* are thinner and more frequent. Two specimens referred by Nikolov and Mandov to *B. moesica* Nik. et Man. have clearly seen fasciculate ribbing (Nikolov and Mandov, 1967, Pl. 1, figs. 2 and 4). Since we limit the genus *Berriasella* to forms with bifurcate and single ribs only, these specimens cannot be assigned to it.

Distribution. Berriasian, *jacobi–boissieri* zones (*paramimounum* subzone) in Bulgaria and France; *jacobi–occitanica* zones in the Crimea.

Material. Four specimens (6–9/13098) from the eastern Crimea (Feodosia region, Sultanovka Village) and the Tonas River basin, collected by Druschits and Bogdanova.

Berriasella subcallisto (Toucas, 1890)
Plate, figs. 10 and 11.

Hoplites calisto var. *subcalisto*: Toucas, 1890, p. 601, Pl. 17, figs. 4 and 5.

Berriasella subcalisto: Mazenot, 1939, p. 53, Pl. 3, figs. 11 and 14 (not figs. 12 and 13); Arnould-Saget, 1953, p. 32, Pl. 3, fig. 9; Nikolov, 1960, p. 159, Pl. 4, figs. 1 and 3; Pl. 5, fig. 4; Dimitrova, 1967, p. 101, Pl. 48, fig. 8; Le Hégarat et Remane, 1968, Pl. 5, fig. 4.

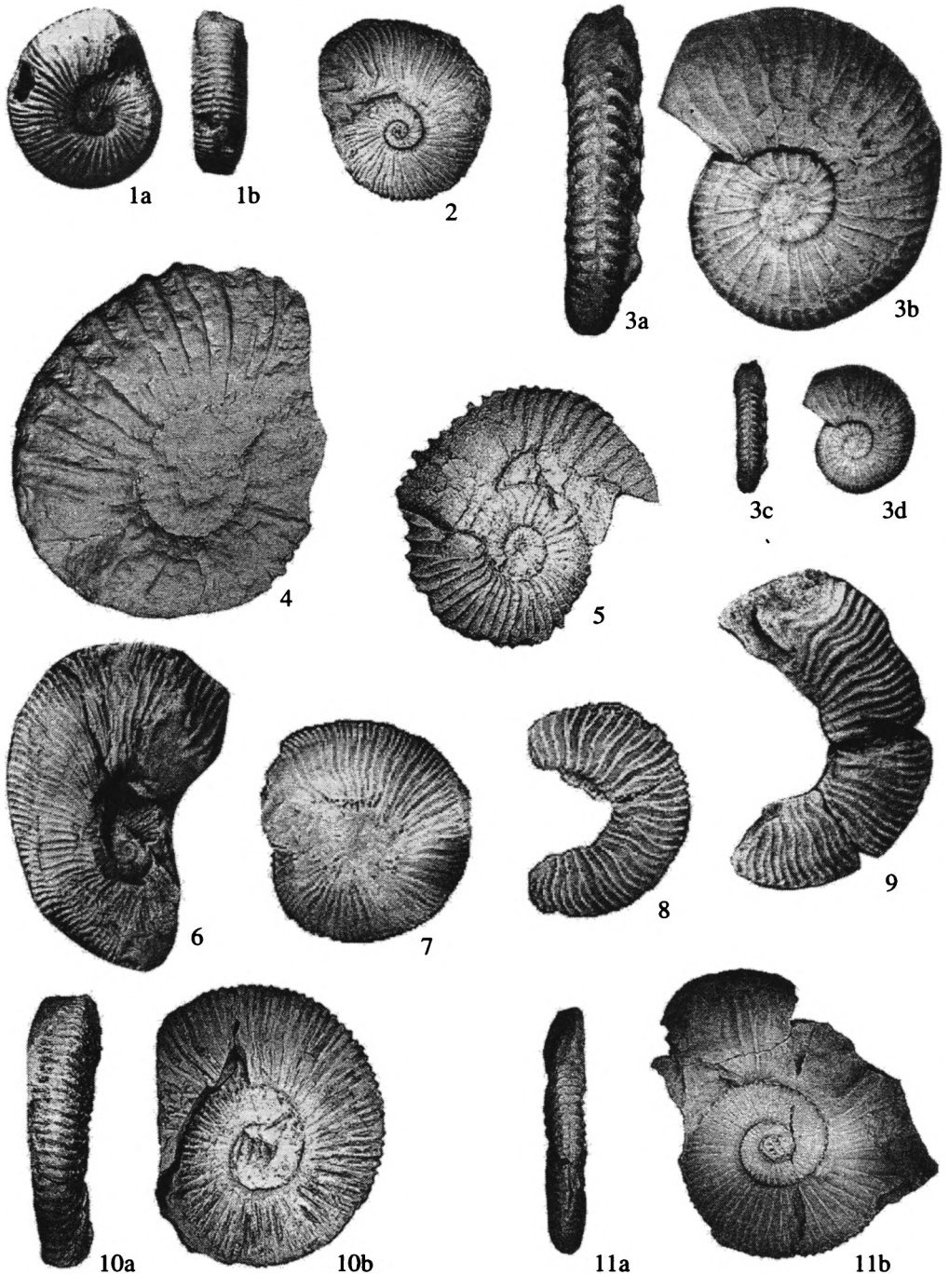


Plate. *Berriasella* species from the Berriasian of the Crimean Mountains.

(1, 2) *Berriasella jacobi* Mazenot: (1) 1/13098, lateral (a) and ventral (b) views; central Crimea, Sary-Su River; Berriasian, *jacobi* Zone, collected by Druschits; (2) 3/13098, lateral view; eastern Crimea, Feodosia region, Barakol'skaya Valley; Berriasian, *jacobi* Zone, collected by Druschits. (3) *Berriasella oppeli* (Kilian), 17/13098, ventral (c) and lateral (d) views, (a, b) the same (×2.5); eastern Crimea, Feodosia region; Berriasian, *jacobi* Zone, collected by Druschits. (4, 5) *Berriasella chomeracensis* (Toucas): (4) 23/13098, ×1.5; (5) 22/13098, lateral view, ×4.0; eastern Crimea, Feodosia, Svyatogo Il'i Cape; Berriasian, *jacobi* Zone, *chomeracensis* Subzone, collected by Arkad'ev and Savel'eva. (6, 7) *Berriasella moesica* Nikolov et Mandov: (6) 7/13098, lateral view; eastern Crimea, Feodosia; Berriasian, *occitanica (?)* Zone, collected by Druschits; (7) 6/13098, lateral view; central Crimea, Tonas River; Berriasian, *jacobi* Zone, collected by Bogdanova. (8, 9) *Berriasella callisto* (d'Orbigny): (8) 20/13098, lateral view; (9) 21/13098, lateral view; central Crimea, Mramornoe Village; Berriasian, *boissieri* Zone, collected by Lysenko. (10, 11) *Berriasella subcallisto* (Toucas): (10) 10/13098, ventral (a) and lateral (b) views; eastern Crimea, Nanikovo Village, Koklyuk; Berriasian, *jacobi* Zone, collected by Druschits; (11) 12/13098, ventral (a) and lateral (b) view; eastern Crimea, Sultanovka (Yuzhnoe) Village; Berriasian, *jacobi* Zone, collected by Nerodenko.

Berriasella (Berriasella) subcallisto: Le Hégarat, 1973, p. 66, Pl. 39, fig. 2 (not Pl. 6, figs. 3–6); Nikolov, 1982, p. 56, Pl. 10, fig. 2; Pl. 11, fig. 1; Tavera, 1985, p. 245, Pl. 34, fig. 3, Fig. 19/M; Liu and Wang, 1987, Pl. 2, fig. 10.

Shape. Shell is of medium size; flanks slightly convex, high; venter narrowed, flat-rounded. Whorl section highly extended, oval in shape. Umbilicus wide, stepped, with low, steeply inclined wall.

Sculpture. Flanks are ornamented with bifurcate and rare single ribs (2 to 3 per whorl). Ribs bifurcate at the middle or within upper one third of the whorl, with posterior branches slightly deflected backward. On the phragmocone, ribs straightly cross flanks, while on the body chamber they are slightly incurved anteriorly. Ornamentation of venter is somewhat different in various specimens. In specimen 10/13098, a narrow furrow interrupting ventral ribs extends almost throughout the adult whorl (D = 45–50 mm). At the end of the whorl, furrow is missing and ribs cross venter at a right angle. In specimen 14/13098 (D = 25–30 mm), siphonal furrow is indistinct, causing the lowering of uninterrupted ribs.

Dimensions (mm) and number of ribs per whorl

Specimen	D	H	W	D_u	H/D	W/D	D_u/D	Number of ribs	
								outer	inner
11/13098	28	10.5	7	10.5	0.38	0.25	0.38	–	45
12/13098	47	17	–	18	0.36	–	0.38	89	46
10/13098	57	18	–	15.5	0.32	–	0.27	80	42

Comparison and remarks. A furrow on venter is an inconsistent character of described ammonites. *B. subcallisto* (Touc.) can be likely considered as a transitional species between *Pseudosubplanites* and *Berriasella* forms. We refer it to the genus *Berriasella* on condition that the forms with trifurcate ribs, for instance, those illustrated by Le Hégarat (1973, Pl. 6, figs. 3–6), are excluded from the genus.

B. subcallisto (Touc.) differs from *B. jacobi* Mazenot by less thick whorls that increase slower, from *B. privasensis* (Pictet) by less coarse and frequent ribbing, and from *B. callisto* (d'Orb.) by a higher point of rib bifurcation.

Distribution. Berriasian, *jacobi–occitanica* (*subalpina* Subzone) zones in the Crimea, Bulgaria, France; *jacobi* Zone in Spain and Himalayas; Berriasian in the Caucasus, Algeria, and Tunisia.

Material. Seven specimens (10–16/13098) from the eastern Crimea (Feodosia region, Sultanovka Village, Barakol'skaya Valley), collected by Druschits, Yanin, Nerodenko, and Bogdanova.

Berriasella oppeli (Kilian, 1889)
Plate, fig. 3.

Ammonites calisto: Zittel, 1868, p. 100, Pl. 20, figs. 1–4 (not fig. 5).

Perisphinctes oppeli: Kilian, 1889, p. 662.

Berriasella oppeli: Mazenot, 1939, p. 49, Pl. 3, figs. 1–3, 5–8 (not fig. 4); Arnould-Saget, 1953, p. 27, Pl. 3, figs. 1, 2, and 4; Nikolov, 1960, p. 160, Pl. 3, figs. 3 and 7; Druschits, 1960, p. 275, Pl. 20, fig. 3 (not fig. 2).

Berriasella (Berriasella) oppeli: Le Hégarat, 1973, p. 58, Pl. 5, figs. 1 and 2; Pl. 38, figs. 4 and 5; not Patrulius and Avram, 1976, p. 172, Pl. 4, figs. 1–4; Nikolov, 1982, p. 52, Pl. 9, figs. 1–6; Tavera, 1985, p. 252, Pl. 35, figs. 3–5, Fig. 19/K; Liu and Wang, 1987, Pl. 1, fig. 8.

Shape. Small microconchs (D up to 23 mm); shell strongly compressed laterally, with almost parallel flanks and narrow rounded venter; whorls oval in cross-section, highly extended; umbilicus wide, stepped, with steeply inclined umbilical wall fluently grading into the flank.

Sculpture. Flanks are decorated with bifurcate and rare single ribs (3–5 per whorl) originating on the umbilical wall. Ribs cross the flank straightly or are slightly incurved. Bifurcation points are at a half or two thirds of the whorl height. On venter, ribs are interrupted by a distinct narrow furrow crossing them at a right angle. The furrow extends throughout the last whorl.

Dimensions (mm) and number of ribs per whorl

Specimen	D	H	W	D_u	H/D	W/D	D_u/D	Number of ribs	
								outer	inner
18/13098	20	7.5	5.2	5.7	0.38	0.28	0.29	77	41
17/13098	23	10.5	4.5	6.5	0.46	0.20	0.28	62	31

Comparison and remarks. *B. oppeli* (Kil.) has ventral furrow well seen particularly on earlier whorls. According to this character, our specimens (microconchs) are closely similar to species figured by Zittel (1868, Pl. 20, fig. 3). Specimens pictured by Patrulius and Avram (1976, Pl. 4, figs. 1–4) are lacking the dis-

tinct ventral furrow, and we do not include them in synonymy of *B. oppeli* (Kil.).

The specimen figured by Druschits (1960, Pl. 20, fig. 2) is small, showing fasciculate ribbing and should be referred most likely to *Pseudosubplanites lorioli* (d'Orb.). *B. oppeli* (Kil.) differs from similar *B. callisto* (d'Orb.) and *B. subcallisto* (Touc.), as they have no ventral furrow.

Distribution. Berriasian, *jacobi–occitanica* zones in the Crimea and Bulgaria; *jacobi* Zone in Spain, southeastern France, and Himalayas; Berriasian in Silesia, Romania, Tunisia, and Argentina (?).

Material. Three specimens (17–19/13098) from the eastern Crimea (Feodosia region, Nanikovo Village, Barakol'skaya Valley), collected by Druschits and Bogdanova.

Berriasella callisto (d'Orbigny, 1847)

Plate, figs. 8 and 9.

Ammonites calisto: d'Orbigny, 1847, p. 551, Pl. 213, figs. 1 and 2; not Zittel, 1868, p. 100, Pl. 20, figs. 1–5.

Ammonites (Perisphinctes) calisto: Favre, 1880, p. 37, Pl. 3, figs 6 and 7 (not fig. 5).

not *Hoplites calisto*: Kilian, 1889, p. 661, Pl. 31, fig. 3; not Toucas, 1890, p. 600, Pl. 17, fig. 3 (= *Berriasella jacobi* Mazenot).

Hoplites calisto: Retowski, 1893, p. 260, Pl. 3, fig. 1a.

Berriasella callisto: Mazenot, 1939, p. 56, Pl. 4, figs. 7–10, 12 (not figs. 6 and 11); Nikolov, 1960, p. 158, Pl. 3, figs. 5 and 6; Pl. 4, fig. 2; Pl. 5, fig. 3; Dimitrova, 1967, p. 101, Pl. 48, fig. 6; Le Hégarat et Remane, 1968, p. 29, Pl. 1, fig. 7 (not Pl. 2, fig. 1).

Berriasella (Berriasella) callisto: Le Hégarat, 1973, p. 53, Pl. 4, figs. 1–3; Pl. 38, fig. 11; Nikolov, 1982, p. 58, Pl. 11, figs. 5 and 6; Pl. 12, figs. 1–5 (not fig. 6); Pl. 13, figs. 1 and 2 (not fig. 3); Tavera, 1985, p. 256, Pl. 35, figs. 8 and 9 (not fig. 7), Fig. 19/N.

Shape. Judging from two broken specimens, shell is evolute, with gently convex flanks. Venter is narrow-rounded; whorl cross-section oval in shape, highly extended; umbilicus wide, with steep wall.

Sculpture. Flanks are ornamented with bifurcate and rare single ribs originating on umbilical wall. Ribs straightly cross flanks, being cranked in the aperture area. Ribs bifurcate at the mid-flank or somewhat lower, with the posterior branch deflected backward. The described specimens lack ventral furrow; ribs cross the venter straightly.

Dimensions (mm) and number of ribs per half a whorl

Specimen	D	H	W	D_u	H/D	W/D	D_u/D	Number of ribs	
								outer	inner
20/13098	41?	15.5	–	–	0.38?	–	–	36	20

Comparison and remarks. In addition to forms with bifurcate and single ribs, researchers (Favre, 1880; Mazenot, 1939; Le Hégarat et Remane, 1968; Nikolov, 1982; Tavera, 1985) referred to *B. callisto* (d'Orb.) ammonites with polygyrate and fasciculate ribbing. We exclude such specimens from the genus *Berriasella* and, correspondingly, from the synonymy of *B. callisto* (d'Orb.). The described species differs from similar *B. subcallisto* (Toucas) by a lower point of rib branching.

Distribution. Berriasian (*boissieri* Zone, *picteti–callisto* subzones) in the Crimea, France, Italy, and Bulgaria; Berriasian in Madagascar; Berriasian (*boissieri* Zone)–Valanginian (*Otopeta* Zone) in Spain.

Material. Two specimens (20–21/13098) from the central Crimea (Mramornoe Village) collected by Lysenko.

Berriasella chomeracensis (Toucas)

Plate, figs. 4 and 5.

Hoplites callisto var. *chomeracensis*: Toucas, 1890, p. 601, Pl. 17, figs. 8 and 9.

Berriasella chomeracensis: Mazenot, 1939, p. 62, Pl. 6, figs. 1–7; Arnould-Saget, 1953, p. 36, Pl. 4, fig. 4, Pl. 6, fig. 5; Nikolov, 1960, p. 161, Pl. 4, figs. 4–6, not Pl. 5, fig. 1; Dimitrova, 1967, p. 102, Pl. 48, fig. 4; Liu and Wang, 1987, Pl. 2, fig. 3.

Berriasella (picteticeras) chomeracensis: Le Hégarat, 1973, p. 70, Pl. 39, fig. 12, not Pl. 7, figs. 3 and 4; Sapunov, 1979, p. 175, Pl. 56, fig. 3; Nikolov, 1982, p. 65, Pl. 14, figs. 1–5.

Berriasella (picteticeras) cf. *chomeracensis*: Patrulius and Avram, 1976, p. 176, Pl. 5, fig. 5.

Berriasella (Berriasella) chomeracensis: Tavera, 1985, p. 244, Pl. 34, fig. 5, Fig. 19/F.

Shape. Shells mostly small, with flattened flanks; umbilicus wide, stepped, with low steep wall.

Sculpture. Flanks are ornamented with simple single and bifurcate ribs. Originating on the umbilical wall, ribs of umbilical shoulder are slightly deflected posteriorly and cross flanks straightly or being incurved a little anteriorly. Nearby the body chamber, ribs are incurved stronger. Specimens with the diameter up to 16–

18 mm are decorated with straight single ribs only. Double ribs appear on shells with D ≈ 20 mm. Ribs bifurcate at the second thirds of the whorl height, and then posterior branches become deflected backward. Single inserted ribs rarely occur. Ribbing is thin and frequent on juvenile whorls in contrast to more coarse and sparse ribs of adult whorls. Ventral ribs are interrupted by a sharp furrow, near which the rib endings are somewhat thickened.

Dimensions (mm) and number of ribs per whorl

Specimen	D	H	W	D_u	H/D	W/D	D_u/D	Number of ribs	
								outer	inner
22/13098	16	6.5	6	–	0.41	0.38	–	30	30
23/13098	22.5	10	7.5	–	0.44	0.33	–	36	32
24/13098	54	20	21	–	0.37	0.39	–	69?	34

Comparison and remarks. In distinction from other *Berriasella* species, *B. chomeracensis* has sharp ribbing and single ribs on initial whorls. The Crimean specimens are most similar to forms described by Tavera (1985) from Spain and by Sapunov (1979) from Bulgaria. In our opinion, single ribs on initial whorls represent a distinctive feature that puts in a doubt the species affiliation the genus *Berriasella.* However, the unbiased comparison between *Berriasella* species is impossible until their complete morphogenesis and undulations of lobe line are known.

Distribution. Berriasian, *jacobi* Zone (*chomeracensis* Subzone) in the Crimea, *jacobi* Zone in southeastern France, Spain, and Himalayas, *jacobi–boissieri* zones in Bulgaria, *chomeracensis* Zone in Tunisia.

Material. Four specimens (22–25/13098) from the Eastern Crimea (Feodosia region, Svyatogo Il'i Cape and Dvuyakornaya Bay) collected by Arkad'ev and Savel'eva.

CONCLUSIONS

Revising collection of *Berriasella* species from Crimea, we identified six species, among which *B. moesica* Nik. et Man. and *B. chomeracensis* (Touc.) are first encountered. *B. jacobi* Maz. is characteristic of the Berriasian lower zone. *B. moesica* Nik. et Man. and *B. subcallisto* (Touc.) are likely indicative of younger Berriasian horizons (the *occitanica* Zone) present in the section, since the former is known in the stratotype area from *jacobi–boissieri* zones and the latter from the *jacobi–occitanica* zones. In southeastern France, *B. oppeli* (Kil.) is characteristic of the *jacobi* Zone and of the *jacobi–occitanica* zones in Bulgaria. *B. callisto* (d'Orb.) implies that *boissieri* Zone is most likely present in Crimean sections, since in France, Spain, Italy, and Bulgaria it is known only from that stratigraphic level, although in Spain it ranges from the Berriasian *boissieri* to the Valanginian *Otopeta* Zone. *B. chomeracensis* (Touc.) is characteristic of the lowermost Berriasian beds or the *chomeracensis* Subzone (Arkad'ev and Savel'eva, 2002).

New data on ammonite distribution in the Feodosia section are consistent with data on ammonoids from Tunisia (Memmi and Salaj, 1975). Based on materials from Tunisia, Tavera *et al.* (1986) distinguished the upward succession of *chomeracensis, pseudogroteanum,* and *grandis* zones in the lower Berriasian (Table). In Tunisia, *B. chomeracensis* is also characteristic of the lower Berriasian section, however it coexists here with other *Berriasella* species (*B. oppeli, B. jacobi*) and *Delphinella* (*D. delphinensis, D. berthei*). In the Crimea, *B. chomeracensis* has not been identified yet separately or in association with other *Berriasella* species. Evidently the *chomeracensis* Zone range in Tunisia is greater than that of *chomeracensis* Subzone in the Crimean Mountains. The problem of the *jacobi* Zone base and, correspondingly, of the Tithonian upper boundary is still open. The described *Berriasella* species imply that Berriasian sequence of the Crimean Mountains includes the standard *jacobi, occitanica,* and *boissieri* zones.

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The Cretaceous Flora of the Kuk-Kaolak Area, Northern Alaska

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Abstract—Among restudied floras from the Kuk-Kaolak area, three taphofloras are defined. The "lower" one from members 1–4 including the Karmak and Kungok floras and the lower subassemblage of the Kuk flora (Smiley, 1966) is dated back to the middle–early late Albian. The "middle" taphoflora from members 5 and 6 comprising the upper Kuk subassemblage after Smiley and Ketik flora is of the late Albian–Cenomanian age, and the "upper" Turonian flora from Member 7 corresponds to the Smiley's Kaolak flora. The taxonomic composition and ages of above taphofloras suggest that Cretaceous floras from the Kuk-Kaolak area developed gradually and successively. Paleobotanic data disprove the former idea of stratigraphic hiatuses in the section (Smiley, 1972a). Cretaceous sedimentation in the Kuk-Kaolak area terminated in the Turonian but not in the Maastrichtian as was previously believed. Members 1, 2–6, and 7 can be considered as lithological analogues of the Tuktu, Chandler, and Seabee formations, respectively, of the Umiat–Chandler and Sagavanirktok areas of northern Alaska. Younger subdivisions known in the last two areas (Prince Creek, Schrader Bluff, and Sagavanirktok formations) have no analogues in the Kuk-Kaolak area. Lithological members 1–6 of the Kuk-Kaolak area are correlative with the Kukpowruk and Corwin formations in the Utukok-Corwin area, where younger Cretaceous sediments are missing. During the Albian–Paleogene time, geological settings changed gradually and irreversibly so that marine sedimentation and sedimentation in general terminated later in easterly areas of the subregion than in western areas of northern Alaska.

Key words: northern Alaska, Cretaceous, Albian, Cenomanian, Turonian, paleobotany, phytostratigraphy.

INTRODUCTION

In northern Alaska (Fig. 1), Cretaceous–Lower Paleogene sediments are widespread north of the Brooks Range over the territory occupied by the Colville sea basin at that time (Sable and Stricker, 1987). Being connected with the Arctic basin on the north, that sea was bordered by young structures of the Brooks Range on the south and southeast. Cretaceous sedimentary rocks of northern Alaska accumulated material eroded from the Brooks Range, where main deformations and uplifting commenced in the terminal Jurassic–initial Cretaceous and were in progress through the Late Cretaceous and, probably, Cenozoic time (Tailleur and Brosge, 1970; Mull, 1979, 1985). The succession of Cretaceous sediments north of the Brooks Range begins with the basal flysch unit that grades upward into a sequence over 3000 m thick, which is composed of shales and argillites in the lower part and of marine shallow-water and deltaic sediments (Nanushuk Group) higher in the section. The entire succession characterizes a gradual filling of the Colville basin, a deep asymmetrical trough located north of the Brooks Range orogenic belt. It is considered as a typical gradational succession from flysch to molasse facies with the Nanushuk Group representing a post-orogenic molasse (Mull, 1979). The Cretaceous volcanism in the Okhotsk–Chukotka volcanogenic belt was probably a source of material for tuffaceous sediments (mainly bentonites) known in northern Alaska (Herman *et al.*, 1996; Kelley *et al.*, 1999).

The plant-bearing sediments of northern Alaska are referred to the Nanushuk (Albian–Cenomanian) and Colville (Turonian–lower Paleocene) groups (Fig. 1) of the asymmetrical North Alaska synclinorium, the axis of which extending along the 70° N subsides at low angle eastward (Fig. 1). In the synclinorium limbs, rocks are deformed into rather simple usually gentle folds. As is noted, the Albian–Lower Paleogene sedimentary succession of northern Alaska is of a classic post-orogenic molasse type. It was deposited in deltaic sedimentation settings and consists of alternating marine (shallow avandelta) and continental (lagoonal, coastal, estuarine, deltaic, and alluvial plain) facies (Mull, 1979; Sable and Stricker, 1987).

Different structures of Nanushuk Group sections in the western (Utukok-Corwin area) and central to eastern parts (Umiat-Chandler and Sagavanirktok areas) of northern Alaska are interpreted as evidencing the simultaneous existence of two large deltaic systems (Fig. 1), the western Corwin and eastern Umiat deltas of the Albian–Cenomanian time (Ahlbrandt, 1979; Ahlbrandt *et al.*, 1979; Sable and Stricker, 1987; Affolter and Stricker, 1987). Later on, probably at the beginning or in the middle of the Late Cretaceous, the Umiat delta was devoured by the larger and rapidly developed Corwin delta (Ahlbrandt *et al.*, 1979;

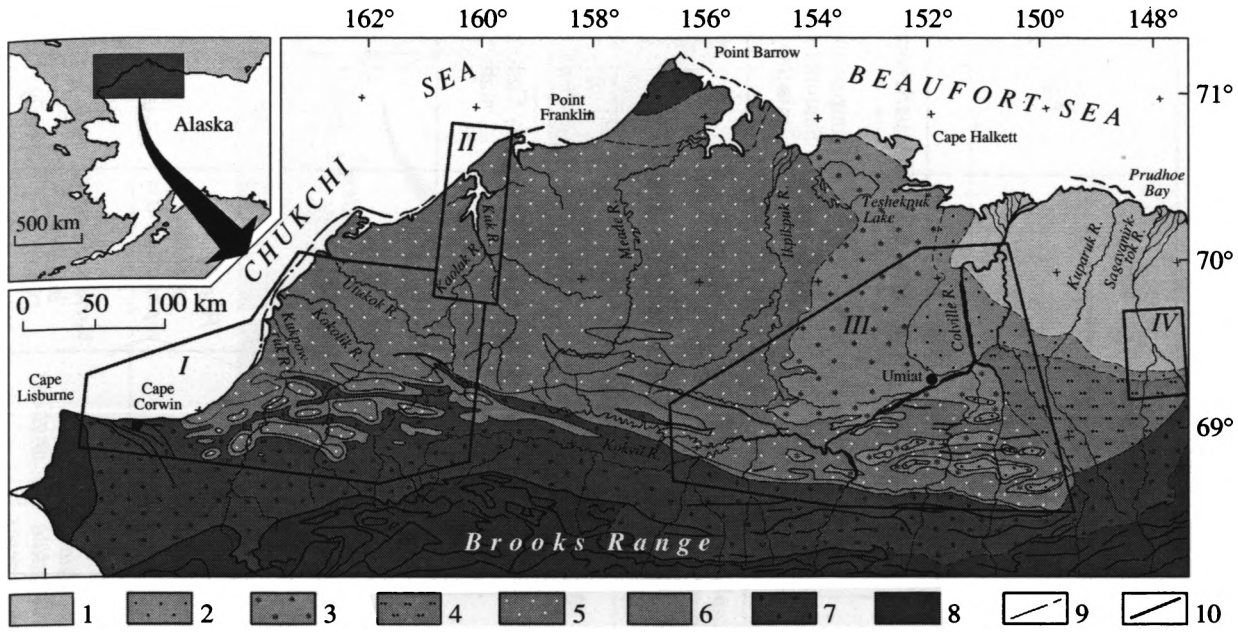

Fig. 1. Schematic geological map of northern Alaska (after Sable and Stricker, 1987, with additions): (1) Tertiary sediments, Sagavanirktok Formation; (2) mostly non-marine and (3) mostly marine sediments of the Colville Group, Upper Cretaceous; (4) Cretaceous sediments, undivided Nanushuk and Colville groups and their analogues; (5) mostly non-marine and (6) mostly marine sediments of the Nanushuk Group, Cretaceous; (7) Lower Cretaceous sediments, mostly the Torok and Fortress Mountain formations; (8) Devonian–Jurassic and older sediments; (9) stratigraphic boundaries; (10) faults. Areas by names: (*I*) Utukok-Corwin. (*II*) Kuk-Kaolak, (*III*) Umiat-Chandler; (*IV*) Sagavanirktok.

Affolter and Stricker, 1987). Traditionally and partly because of structural and compositional differences, the Nanushuk Group sections are subdivided into different stratigraphic units in the western and central to eastern parts of northern Alaska. Correlation between these units is based on marine fossils (Imlay, 1961; Jones and Gryc, 1960; and others) and paleobotanic data.

The Cretaceous flora of northern Alaska was studied since 1956, when Ch.J. Smiley sampled during several field expeditions a large collection of fossil plants (approximately 10000 specimens) from over 250 localities within the area from the Cape Lisburne in the west to the Colville River basin in the east (Smiley, 1966, 1967, 1969a, 1969b, 1972a, 1972b; Scott and Smiley, 1979). He proposed a phytostratigraphic scheme for the lower Albian to Maastrichtian stratigraphic interval (Table 1). Based on sequences of alternating marine and continental plant-bearing sediments, which also replace each other along the strike in some areas, the distinguished "floral zones" were correlated to the marine scale (Chapman and Sable, 1960; Chapman *et al.*, 1964; Brosge and Whittington, 1966; Detterman *et al.*, 1963; Imlay, 1961; Jones and Gryc, 1960; Table 1). Unfortunately, Smiley characterized his scheme briefly, without complete lists of taphofloras characteristic of concrete localities, and only some fossil plants were described and illustrated. The scheme is therefore of a limited practical value. The Smiley's phytostratigraphic scheme includes seven "floral zones" (Table 1)

Floral Zone I (Aptian? to lower–middle Albian) is characterized by abundant and diverse ferns, ginkgoaleans with dissected and linear leaves, and cycadophytes. Conifers are rare except for the *Podozamites* genus, angiosperms are absent.

Floral Zone II (first half of the upper Albian) is characterized by diverse ferns, ginkgoaleans, cycadophytes, and conifers. Angiosperms are not found, except for scarce fragments of probably aquatic plants.

Floral Zone III (second half of the upper Albian to basal Cenomanian) is characterized by significant reduction in abundance of ferns, ginkgoaleans, and cycadophytes, higher diversity of conifers, and by the first appearance of angiosperms. Taphofloras of this zone include last cycadophyte species of northern Alaska.

Floral Zone IV (lower Cenomanian) is characterized by dominant dicotyledons associated with rare conifers and, locally, with abundant. *Ginkgo* leaves. Rare ferns are represented by fragmented fronds, cycadophytes are absent.

Floral Zone V (Turonian) is characterized by abundant and diverse dicotyledons and conifers. Ferns are rare, *Ginkgo* representatives are infrequent, and cycadophytes are absent. Taphofloras from this zone include last representatives of the *Podozamites* genus.

Floral Zone VI (Coniacian? Santonian–Campanian) is characterized by diverse dicotyledons and conifers. Ferns are rare, *Ginkgo* leaves are locally numer-

Table 1. Biostratigraphic correlation of marine and continental sediments of northern Alaska (after Smiley, 1972, with modifications); Arabian numbers designate lithological units in the Kuk–Kaolak rivers area

Floral zone		Lithostratigraphic units: west							Marine faunal zonation			Age
		Corwin Bluff Kukpowruk River		Kuk River	Chandler–Colville rivers				ammonites	inoceramids	foraminifers	
No.	Principal taxa	non-marine sediments	marine sediments	non-marine sediments	non-marine sediments		marine sediments					
VII	Dicotyledons, conifers			7	Kogosukruk Tongue	Prince Creek Formation	Schrader Bluff Formation	Colville Group		*Inoceramus patootensis* *I. steenstrupi*	*Anomalinoides talaria* *Nonionella asutinana*	Maastrichtian? / Campanian–Santonian (Coniacian?)
VI	Dicotyledons, conifers				Tuluvak Tongue							
V	Dicotyledons, conifers			6			Seabee Formation		*Scaphites delicatulus* *Watinoceras* *Borissjacoceras*	*I. labiatus* *I. cuvieri*	*Trochammina ribstonensis* *Gaudryina irinensis*	Turonian
Regional stratigraphic unconformity												
IV	Conifers, dicotyledons				Niakogon Tongue	Chandler Formation	Ninuluk Formation	Nanushuk Group		*I. dunveganensis*	*Trochammina rutherfordi* *Gaudryina canadensis*	early Cenomanian
III	All principal taxa	Corwin Formation	Kukpowruk Formation	5	Killik Tongue, upper part		Grandstand Formation		*Cleoniceras* *Gastroplites* *Paragastroplites*	I. cadottensis I. anglicus	*Verneuilinoides borealis*	late
II	Ferns, ginkgoaleans, cycadophytes, conifers			4								
				2.3	Killik Tongue, lower part							
IB	Ferns, ginkgoaleans, cycadophytes			1			Tuktu Formation				*Haplophragmoides topagorukensis*	middle
							Torok and Fortress Mountain Formations		*Colvillia crassicostata* *Beudanticeras*	*I. anglicus*	*Verneuilinoides borealis* *Dorothea chandlerensis*	early (Aptian?)
IA	Ferns, ginkgoaleans, cycadophytes		Torok and Fortress Mountain Fms.									

Albian

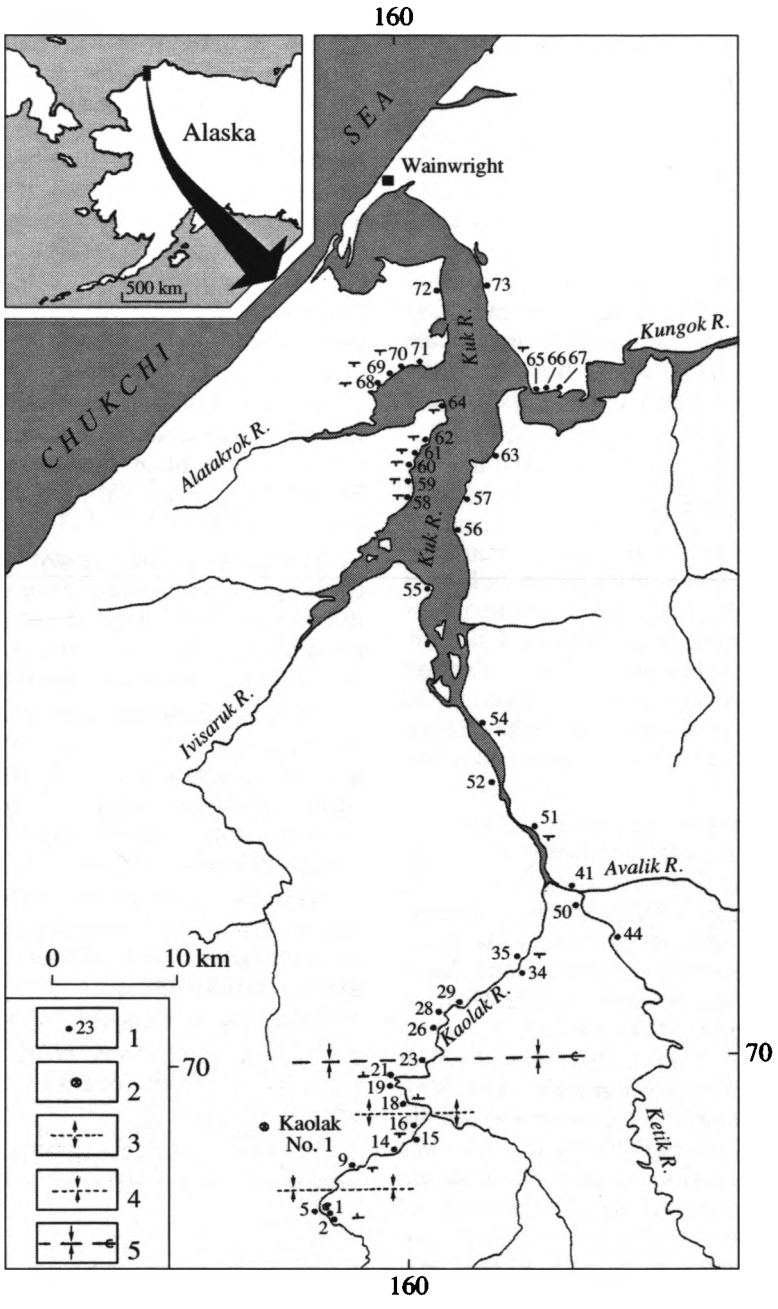

Fig. 2. Localities of fossil plants and elements of geological structure in the Kuk and Kaolak river basins (after Smiley, 1966 with modifications): (1) localities of fossil plants; (2) test borehole; (3) anticline axis; (4) syncline axis; (5) synclynorium axis.

ous. Last ginkgoaleans of northern Alaska occur in taphofloras from this zone.

Floral Zone VII (presumably Maastrichtian or upper Campanian–Maastrichtian) is characterized by diverse angiosperms usually of poor preservation and presence of conifers, particularly from the *Parataxodium* genus. Ferns are rare, ginkgoaleans, cycadophytes, and *Podozamites* representatives are absent.

This paper is dedicated to restudied fossil floras from the Kuk-Kaolak area located in the northwestern part of northern Alaska. At the beginning of the 1950s, J.S. Lowter sampled fossil plants from the Kuk, Kaolak, and Avalik river basins in the Kuk-Kaolak area. Smiley was first to study systematically the Cretaceous floras from the southern part of the Kuk River basin and, particularly, along the Kaolak River (Fig. 2) since 1956. The work aimed at collecting and dating fossil amber was first headed by R.L. Langenheim (Langenheim *et al.*, 1960). Subsequently, this project oriented toward investigation of fossil flora was continued in

1961, when the research involved a territory extending up to the Wainwright Settlement in the north part of the area under consideration (Smiley, 1966). Ch.J. Smiley, H.F. Shorn, and their colleagues collected fossil plants from over 40 localities along the Kuk, Kaolak, Ketik, Alatakrok, and Kungok rivers (Fig. 2). Unfortunately, Smiley described this and other collections sampled in northern Alaska without necessary details. He did not publish complete lists and images of fossil plants from studied localities, and many plants were determined only at generic level. After 1961, no targeted sampling of fossil plants was carried out in this area and all the published very incomplete data on the flora from the Kuk-Kaolak area can be found only in two papers mentioned above.

MATERIALS AND METHODS

Characterization of floral assemblages from the Kuk-Kaolak area of northern Alaska given below is based on the restudy (together with R.A. Spicer) of fossil plants collected by Smiley in the Kuk and Kaolak river basins in 1956 and 1961 (Fig. 2). This collection of several hundreds of plant impressions is stored in the Smithsonian Institution, Washington DC (the United States). Unfortunately, the collection itself and some samples are not numbered.

The examined plant remains are largely represented by leaf impressions. There are also remains of shoots with leaves, fossil reproductive structures, and wood. Plant remains are frequently well preserved, suitable for examination of fine structural details of plant organs, although phytoleims are usually absent. Samples were photographed under oblique illumination using the small-format camera Nikon F-801 S, macro lense MicroNikkor 2.8/60, and high-resolution film Ilford FP-4. Black-and-white photographs sized 20 × 25 cm were then printed. Leaf drawings were made by Indian ink directly on photographs, which then were washed off by aqueous solution of potassium iodide and iodine. The obtained photographs and drawings of fossil leaves were scanned and used in subsequent analysis. We (Spicer and myself) prepared also the relational database for the Cretaceous Smiley's collection from northern Alaska. The program FileMaker Pro was used to combine the text (stratigraphic and geographic position of the plant remain, taxonomic affinity and name of the plant, sedimentological context of the locality, collection data) and graphic (photograph and line drawing of the plant remain, geological schematic map and stratigraphic column of the locality) information. The database prepared includes approximately 1700 records.

CRETACEOUS STRATIGRAPHY OF THE KUK-KAOLAK AREA

Sediments in the Kuk-Kaolak area occur almost horizontally with dip angles up to 1° and lower. They are confined to northern and southern limbs of synclinorium, the axis of which extends approximately along the latitude of 70° N. The southern limb is complicated by small second-order folds (Fig. 2). The sediments, mostly shales and sandstones with interbeds of conglomerates, siltstones, and bentonites, yield numerous plant remains: leaf impressions, shoots with leaves, and wood fragments. Plant remains are well preserved and usually confined to siderite concretions. The rocks enclose coal seams up to 30 cm thick occurring throughout the entire section; thicker (up to 2.5 m) coal beds are rare in the lower part of the section and more abundant in its upper part.

Seven lithological members are defined in the Kuk-Kaolak Cretaceous section exposed along creeks and rivers, where plant remains were sampled (Fig. 2). From the base upward, these are the following units (Smiley, 1966):

Member 1. Dark gray shales, locally silty and coaly; plant remains are found only in the upper part of the member (localities 72 and 73) approximately 200 m thick and exposed up to the ocean coast. The thickness of the upper part of the member is up to 1300 m.

Member 2. Sandstones from red to gray, with ripple marks and cross bedding, locally with well rounded pebbles. There are cavities left by decomposed wood fragments and stems, which are filled with sediments; plant remains are from localities 71, 72. The thickness is approximately 40 m.

Member 3. Shales and silty shales with thin sandstone interbeds and coal seams up to 2.7 m thick; plant remains are from localities 59–71. The thickness is approximately 230 m.

Member 4. Coarsely alternating shales and compact hard sandstones with thin coal seams; plant remains are from localities 55–58. The thickness is approximately 130 m.

Further, sediments approximately 200 m thick are unexposed or poorly exposed; plant remains occur in localities 54–55.

Member 5. Shales and silty shales with rare interbeds of compact sandstones, thin coal seams, bentonite clays, and rare amber inclusions; plant remains are from localities 41–54. The thickness is approximately 220 m.

Member 6. Thickly alternating shales and sandstones with thin coal seams, white bentonites, bentonitic clays, and rare amber inclusions; plant remains are sampled from localities 1–16 and 29–41. The thickness is approximately 165 m.

Member 7. Alternating sandstones, sandy clays, coals, white bentonites, and bentonitic clays, amber inclusions; plant remains are from localities 18–28. The thickness is approximately 85 m.

A similar section represented by members 1–5 and the lower part of Member 6 (Smiley, 1966) is penetrated by the test borehole 1 "Kaolak" (Collins, 1958)

drilled in the southern part of the area (Fig. 2). The poorly recovered interval of the borehole section between members 4 and 5 is composed of alternating sandstones, shales and coal seams. In the upper 35-m-thick interval, the recovered borehole section is composed of loose sands and gravel referred to the Pleistocene Gubik Formation.

The section of the Kuk-Kaolak area is lacking signs of the regional stratigraphic and angular unconformity, which was established between the Ninuluk–Niakogon and overlying Seabee formations in the easterly located Umiat-Chandler area (Detterman *et al.*, 1963). The geological survey of the area under consideration failed to reveal this unconformity as well (Smiley, 1966).

CORRELATION AND AGE OF TAPHOFLORAS FROM THE KUK-KAOLAK AREA

Smiley (1966) defined five plant assemblages in the section mentioned above and named them (from older to younger) as the Karmak, Kungok, Kuk, Ketik, and Kaolak floras. Their characteristics presented below are based on data by Smiley with addition of my results obtained after the collections restudy at the Smithsonian Institution.

The *Karmak flora* is from the upper part of lithological Member 1 with rare plant remains, localities 72 and 73 (Fig. 2). Smiley identified four genera *Baiera, Zamites, Amentotaxus*, and *Podozamies* in this flora, and I determined only two plant species from these localities (Table 2). The age of the spore–pollen assemblage from Smiley's locality 72 in the Kuk River mouth area (Fig. 2) was determined by Stanley (1967) as the Albian, based on the absence of gymnosperm pollen and presence of spores and pollen identical or similar to those from the uppermost Lower Cretaceous sediments of Siberia and Canada. A similar spore–pollen assemblage, which is also dated back to the Albian (Stanley, 1967), is described from approximately coeval sediments in the lower reaches of the Meade River (Fig. 1).

The *Kungok flora* is from the upper part of lithological Member 3, localities 59–71 (Fig. 2). According to Smiley, it consists of several fern species (*Dicksonia* spp. included), abundant ginkgoaleans represented by deeply dissected leaves of *Baiera* and *Ginkgo*, frequent cycadophytes *Ctenis, Nilssonia*, and *Podozamites*, conifers *Podozamites, Elatides (?Araucarites), Pagiophyllum (?Abietites), Amentotaxux, ?Juniperus, Cephalotaxopsis*, Taxodiaceae resembling *Taiwania*, and first scarce angiosperm forms represented by two impressions of small *Nelumbites* leaves. My determinations of plants from these localities are given in Table 2.

The *Kuk* flora is described from lithological members 4 and 5, localities 51–58 (Fig. 2). In Smiley's opinion, this flora is characterized by the presence of ferns and gymnosperms (ginkgoaleans with dissected leaves included) and differs from older floras from the area under consideration in rare occurrence or absence of cycadophytes, *Pagiophyllum (?Abieitites)* and conifers similar to *Taiwania*, in the appearance of *Sequoia, Cephalotaxus, Torreya*, and rare terrestrial angiosperms (woody dicotyledons) that appear in lithological Member 5 (localities 51, 52, and 54). Smiley noted its transitional composition between the older (Karmak and Kungok) and younger (Ketik and Kaolak) floras and distinguished two the Lower Kuk subassemblage close to older floras (from lithological Member 4) and the Upper Kuk subassemblage (from Member 5) with angiosperms, *Sequoya*, and other "recent" conifers. My determinations of plants from members 4 (Lower Kuk flora) and 5 (Upper Kuk flora) are given in Table 2.

The *Ketik flora* is confined to the upper part of lithological Member 6, localities 29–41 and 1–16 (Fig. 2). Smiley noted that characteristic of the flora are angiosperms locally abundant and diverse, the frequent *Platanophyllum* and *Credneria* included. Locality 41 yielded abundant impressions of leaves resembling those of birch and oak (in my opinion, he took lobed leaflets of compound *Dalembia* leaves for the oak leaves), which occur together with *Cissites, Menispermites*, and *Populites*. Younger layers (locality 44) enclose first *?Juglans* and *Viburnum* along with rare entire-leaved *Celastrophylum, Dalbergites, Laurophyllum*, and *?Magnolia*. Ferns are rare, while cycadophytes and ginkgoaleans with dissected leaves are missing; forms of the last group resemble recent *Ginkgo biloba* L. Conifers are represented by recent genera *Sequoia* (shoots with leaves and cones), *?Taxodium* (shoots with leaves), *Sequoiadendron* (cones), *Juniperus* (shoots with leaves), *Cephalotaxus* (shoots with leaves and cones), *?Torreya* (shoots with leaves and cones), *Pinus* (needles, cone scales, pollen), *Picea* (cones, pollen), and *?Larix* (clusters of needles), and by primitive genera *Podozamites* and *Cephalotaxopsis*. The *Amenotaxux* genus common in older floras is missing. My determinations of plants from these localities are given in Table 2.

The *Kaolak flora* is from the upper part of lithological Member 7, localities 18–28 (Fig. 2). According to Smiley, this flora is dominated by large-leaved Platanaceae with common *Trochodendroides* and *Menispermites*. Entire-margined angiosperms are represented by a single leaf with a thick texture, the taxonomic affinity of which is unclear. Ferns are rare, although spore–pollen spectra contain abundant fern spores of several types (Langenheim *et al.*, 1960). Ginkgoaleans, cycadophytes, primitive conifers, and representatives of the genera *Podozamites, Cephalotaxus*, and *Juniperus* are missing. Angiosperms are represented by the recent conifer genera *Sequoia, Sequoiadendron, ?Taxodium*, and *?Torreya*. Characteristic of spore–pollen spectra are abundant conifers dominated by Taxodiaceae (Langenheim *et al.*, 1960). Based on the flora composition, Smiley assumed that conifers prevailed in the elevated areas of northern Alaska during the time under consideration, whereas lowlands were occupied by

Table 2. Composition of fossil plant assemblages from the Kak-Kaolak area (flora names and numbers of localities after Smiley, 1966)

Species \ Flora, locality	Karmak flora, localities 72, 73	Kungok flora, localities 59–71	Lower Kuk flora, localities 55–58	Upper Kuk flora, localities 51–54	Ketik flora, localities 1–6, 29–41, 44, 50	Kaolak flora, localities 18–28
	1	2	3	4	5	6
Thallites sp.		+				
Equisetites ex gr. *bureyensis* (Heer) Krysht.			+			
Equisetites sp.		+				
Sphenopteris sp.			+	+		
Stenorachis striolatus (Heer pars) Nathorst				+		
Onychiopsis elongata (Geyl.) Yok.		+	+	+		
Onychiopsis cf. *elongata* (Geyl.) Yok.						+
Onychiopsis psilotoides (Stokes et Webb) Ward		+				
Gleichenites cf. *asiatica* Philippova				+	+	
Gleichenites sp.					+	
Coniopteris sp.		+	+	+		
Birisia cf. *alata* (Prynada) Samylina		+	+		+	
Birisia ochotica Samylina						+
Birisia sp.			+			
Arctopteris penzhinensis E. Lebedev				+		
Arctopteris cf. *rarinervis* Samylina		+				
Arctopteris sp.				+		
Dennstaedtia sp.				+		
Taeniopteris ex gr. *lundgreni* Nathorst				+		
Taeniopteris sp.		+				
Nilssonia cf. *decursiva* Samylina		+				
Nilssonia cf. *magnifolia* Samylina		+				
Nilssonia menneri E. Lebedev		+				
Nilssonia yukonensis Hollick				+		
Heilungia sp.		+				
Ctenis sp.		+				
Ginkgo ex gr. *adiantoides* (Unger) Heer		+		+	+	
Ginkgo ex gr. *concinna* Heer				+		
Ginkgo ex gr. *huttoni* (Sternb.) Heer		+				
Ginkgo ex gr. *lepida* Heer		+				
Sphenobaiera cf. *czekanowskiana* (Heer) Florin		+				
Sphenobaiera ex gr. *czekanowskiana* (Heer) Florin				+		
Sphenobaiera sp.		+		+		
Sphenobaiera (?) sp.			+		+	
Phoenicopsis ex gr. *angustifolia* Heer				+		
Desmiophyllum magnum (Samyl.) Samyl.	+	+	+			
Desmiophyllum sp.		+	+	+	+	
Podozamites ex gr. *eichwaldii* Schimper				+		
Podozamites ex gr. *lanceolatus* (Lindl. et Hutton) Braun		+			+	
Podozamites spp. (isolated leaves)		+	+		+	
Pagiophyllum triangulare Prynada		+				
Pagiophyllum sp.		+		+	+	
Sequoia tenuifolia (Schmalh.) Sveshn. et Budantsev						+
Athrotaxopsis ex gr. *expansa* Fontaine		+	+	+	+	

Table 2. (Contd.)

Species	1	2	3	4	5	6
Elatochadus smittiana (Heer) Seward		+				
Elatocladus sp.					+	
Pityophyllum ex gr. *nordenskioldii* (Heer) Nath.	+	+	+	+		
Pityophyllum ex gr. *staratschinii* (Heer) Nath.			+	+	+	+
Abietites sp.		+				
Araucarites anadyrensis Krysht.		+	+	+		
"*Araucarites*" sp. (cone)				+		
Cephalotaxopsis intermeida Hollick				+	+	+
Carpolithes spp.		+		+	+	
Conifer cones (several types)		+		+	+	+
"*Magnolia*" cf. *lakesii* (Knowlton)					+	
Nelumbites sp.		+				
Menispermites kryshtofovichii Vachrameev					+	
Menispermites cf. *reniformis* Dawson					+	
Menispermites cf. *septentrionalis* Hollick					+	
Menispermites ex gr. *septentrionalis* Hollick						+
Platanus cf. *latiloba* Newberry				+		
Platanus cf. *louravetlanica* Herman				+	+	
Platanus cf. *primaeva* Lesquereux						+
Arthollia inordinata (Hollick) Herman					+	
Arthollia cf. *pacifica* Herman						+
"*Credneria*" cf. *grewiopsoides* Hollick						+
Pseudoprotophyllum boreale (Dawson) Hollick						+
Pseudoprotophyllum sp.						+
Platanaceae gen. et sp. indet.					+	+
Populites cf. *pseudolancastriensis* Hollick					+	
Populites sp.						+
"*Lindera*" *media* Philippova					+	
Schefferaephyllum venustum (Philip.) Philip.					+	
Celastrophyllum sp.					+	
Araliaephyllum (?) sp.				+		
Leguminosites sp.						+
Dalbergites cf. *simplex* (Newberry) Seward					+	
Dalbergites sp.					+	
Viburniphyllum (?) sp.					+	
Grebenkia anadyrensis (Krysht.) E. Lebedev					+	
Cissites sp.					+	
Zizyphoides sp.						+
Trochodendroides ex gr. *arctica* (Heer) Berry						+
Trochodendroides notabilis Herman						+
Trochodendroides sp.						+
Dalembia vachrameevii E. Lebed. et Herman					+	
Dicotylophyllum spp.				+	+	+
Fossil wood						+
Root remains		+				

angiosperm plants. My determinations of plants from these localities are given in Table 2.

Discussing the age of all the floras mentioned above, Smiley (1966) correlated the Kuk-Kaolak section with sections of the Utokok-Corwin and Umiat-Chandler areas taking into consideration the distribution of bentonites in all the sections and taxonomic composition of relevant floras. He correlates lithological Member 1 of marine gray shales with the marine Tuktu Formation of the Albian age (Table 1), because core samples from the member upper part penetrated by the Kaolak test borehole yielded *Entolium* sp. indicating, according to Imley (1961, cited after Smiley, 1966), the middle Albian age of host sediments. Thus, the Karmak flora is dated back to the terminal middle Albian.

According to generalization by Smiley (1966), bentonites first appear in the middle Albian Topagoruk Formation in sections of the Umiat-Chandler area (Colville River basin), are rare in the Albian–Cenomanian Nanushuk Group, and abundant in overlying sediments of the Colville Group (Prince Creek Formation). In the Kuk-Kaolak area, oldest bentonite beds occur at the base of lithologic Member 5 (locality 54). They are rare in this member, being abundant in overlying members 6 and 7. Based on these observations, Smiley concluded that lithological members 4 and 5 bearing the Kuk flora are Cenomanian in age and correlative with the upper part of the Killik Tongue, while members 6 and 7 with the Ketik and Kaolak taphofloras are post-Cenomanian, probably Turonian and Coniacian, respectively. Member 6 was correlated with the Prince Creek Formation, an equivalent of the marine Seabee Formation, and Member 7 with the Tuluvak Tongue of the Prince Creek Formation. According to Smiley, Member 3 should be correlated with the lower part of the Killik Tongue and dated, along with the Kungok flora, back to the late Albian.

Although, as is noted above, there are no signs of regional unconformity in the Kuk-Kaolak sections, a sharp transition from the Kuk to the Ketik flora of different taxonomic composition implies, in opinion of Smiley (1966), a hiatus between members 5 and 6. The hiatus was thought to be spanning stratigraphic interval corresponding to the Ninuluk Formation in the Umiat-Chandler area, i.e., to the Niakagon Tongue (upper Cenomanian) and, probably, lower part of the Killik Tongue (Cenomanian).

Subsequently, Smiley (1972a) changed his views on ages of floras from the Kuk-Kaolak area (Table 1). As before, the Karmak and Kungok floras were dated back to the middle and late Albian, respectively, but lithological members 4 and 5 with the Kuk flora were attributed to the upper Albian (except for its uppermost part). He correlated Member 6 with the Ketik flora, as before, with the Turonian and simultaneously considered Member 7 and the relevant Kaolak flora as the late Campanian–?early Maastrichtian in age thus correlating them with the Kogosukruk Tongue and enclosed flora (Smiley, 1972a), although without any argumentation. Thus, Smiley postulated two considerable hiatuses separating members 5, 6 and 7 of the studied section, which correspond to the uppermost Albian–Cenomanian and Coniacian–early Campanian, respectively.

I think that some of age determinations proposed by Smiley (1966, 1972a) are invalid and may be corrected by correlating the floras with their well-dated counterparts from northeastern Russia and northern Alaska. This primarily concerns the Ketik and Kaolak floras, which are, in my opinion, substantially different in age from assessments proposed by Smiley (1972a). In addition, I think that he overstated the number of floras (five) distinguishable in the Kuk-Kaolak area. In my opinion, only three taphofloras readily correlative with floras from northeastern Asia and other areas of northern Alaska are recognizable here.

The *"lower" taphoflora* from lithological members 1–4 unites the Karmak and Kungok floras and the lower plant subassemblage of the Kuk flora after Smiley (1966). I think that it is unjustified to rank scarce plant remains from uppermost Member 1 as the individual Karmak flora (Smiley, 1966) particularly taking into consideration that the same species are present in the Kungok flora. The lower subassemblage of the Kuk flora is similar to older floras of the area and, thus, should be included into the "lower" taphoflora. The taphoflora consists of 37 species of liverworts (?), horsetails, ferns, cycadophytes, ginkgoaleans, conifers, and angiosperms (Table 2). Characteristic of it is a high diversity of ferns belonging to five genera and cycadophytes represented by four genera, among which *Nilssonia* forms are most diverse (three species). Ginkgoaleans are represented by *Ginkgo* and *Sphenobaiera* genera. *Desmiophyllum magnum*, a species with large elongated arcuate leaves with parallel venation is among most remarkable forms of this taphoflora. Conifers are dominated by isolated leaves of the genus *Podozamites*, which occur in association with subordinate *Pagiophyllum, Athrotaxopsis, Elatocladus, Pityophyllum, Abietites*, and *Araucarites* forms and with cones of several types. Angiosperms are scarce, represented by two impressions of small *Nelumbites* leaves from lithological Member 3 (locality 65, Fig. 2).

The taphoflora in question is undoubtedly similar to the Kukpowruk flora from the Kukpowruk River basin of the Utukok-Corwin area (Fig. 3) (Spicer and Herman, 2001; Herman and Spicer, 2002). In both floras, species in common are *Equisetites* ex gr. *bureyensis, Onychiopsis psilotoides, Birisia* cf. *alata, Arctopteris* cf. *rarinervis, Nilssonia* cf. *decursiva, N.* cf. *magnifolia, N. menneri, Ginkgo* ex gr. *adiantoides, Sphenobaiera* ex gr. *czekanovskiana, Desmiophyllum magnum, Podozamites* spp. (isolated leaves), *Pagiophyllum triangulare, Pityophyllum* ex gr. *nordenskioldii*, and *P.* ex gr. *staratschinii*. Both taphofloras are also marked by the first appearance of rare angiosperms. The Kukpowruk taphofora is dated back to the early(?) late

Albian, the second half of the substage excuded (Spicer and Herman, 2001; Herman and Spicer, 2002). In northern Alaska, the "lower taphoflora" is close to the fossil flora from the lower part of the Killik Tongue of the Chandler Formation (Smiley, 1969a; Fig. 3).

The abundant and diverse Buor-Kemuss flora of the early middle Albian of northeastern Russia is another one close in composition to the "lower" taphoflora from the Kuk-Kaolak area. In Russia, that flora is known from the following rich localities: the Buor-Kemuss Formation, the Zyryanka coal field of the Kolyma River basin (Prinada, 1938; Samylina, 1960, 1964, 1967, 1974); the upper Ainakhkurgen Subformation and the Chimchememel' Formation, the Ainakhkurgen Trough of the Bol'shoi Anyui River basin (Filippova, 1975; Samylina, 1974, 1976); and the Omsukchan and Toptan formations, the Omsukchan area of the Magadan region (Samylina, 1976). Practically all the fossil plants constituting the "lower" taphoflora from the Kuk-Kaolak area are characteristic of taphofloras of the Buor-Kemuss type.

Consequently, the "lower" taphoflora from the Kuk-Kaolak can be dated back to the middle–late Albian (the second half of the last age excluded). On the one hand, it includes *Entolium* sp. of the middle Albian age found in the upper part of lithological Member 1, and is similar to the Kukpowruk [early (?)–late Albian except for the second half of the last age] and Buor-Kemuss (early–middle Albian) floras, on the other hand (Fig. 3). The inferred age is consistent with that proposed by Smiley (1972a) for the floras from lithological members 1–4. In his earlier publication (Smiley, 1966), he referred these members to the middle Albian–Cenomanian.

The *"middle" taphophlora* from lithological members 5 and 6 includes the upper subassemblage of the Kuk flora along with Ketik flora (Smiley, 1966). I consider these two components a single flora, because they are taxonomically similar at the species (taxa in common are *Gleichenites* cf. *asiatica, Ginkgo* ex gr. *adiantoides, Athrotaxopsis* ex gr. *expansa, Pityophyllum* ex gr. *staratschinii, Cephalotaxopsis intermedia, Platanus* cf. *louravetlanica*) and generic (*Sphenobaiera, Desmiophyllum, Podozamites, Pagiophyllum, Dicotylophyllum*) levels. In addition, characteristic of both is a notable share of angiosperms, the abundance and diversity of which is substantially higher even in lithological Member 5 ("the lower Kuk flora") as compared to older floras of the area.

The "middle" taphoflora includes approximately 50 species of ferns, cycadophytes, ginkgoaleans, czekanowskialeans, conifers, and angiosperms (Table 2). Ferns are quite diverse in this taphoflora, and their most interesting representatives are *Gleichenites* cf. *asiatica* and *Arctopteris penzhinensis* missing from the older taphoflora. Cycadophytes are relatively rare, represented by genera *Taenopteris* and *Nilssonia*; in Member 6, they are absent. Ginkgoaleans include genera *Ginkgo* and *Sphenobaiera*, while czekanowskialeans are represented by rare finds of *Phocnicopsis* forms. Leaves with parallel venation, taxonomic affinity of which is unclear, are referred to the genus *Desmiophyllum*. The diverse conifers occur as isolated leaves, shoots with leaves attached, seeds, and cones. Most remarkable in this group are abundant leaves of the genus *Podozamites* and remains of *Araucarites anadyrensis* shoots with hook-like leaves. The high diversity and abundance of angiosperms in lithological member 5 increases further upward in Member 6 (Table 2). The remarkable angiosperm taxa are the entire-leaved forms (genera *Magnoliaephyllum, "Lindera," Schefferaephyillum, Celastrophyllum, Araliaephyllum, Dalbergites*), Platanaceae, frequently with large leaves (genera *"Platanus"* and *Arthollia*), and representatives of the genus *Minispermites* (three species). *Grebenkia anadyrensis* and *Dalembia vachrameevii* are the other species of interest.

The "middle" taphoflora of the Kuk-Kaolak area is similar to the Corwin taphoflora of the late Albian–Cenomanian (Fig. 3) from the Kukpowruk River basin of the Utukok-Corwin area (Spicer and Herman, 2001; Herman and Spicer, 2002), as they both include species (*Ginkgo* ex gr. *adiantoides, G.* ex gr. *concinna, Pityophyllum* ex gr. *nordenskioldii, Araucarites anadyrensis, Cephalotxopsis intermedia, Grebenkia anadyrensis, Schefferaephyllum venustum, Platanus* cf. *louravetlanica*) and genera (*Gleichenites, Coniopteris, Birisia, Arctopteris, Sphenobaiera, Desmiophyllum, Podozamites, Menispermites, Araliephyllum*) in common. In both areas, these are the oldest floras with prevalent and diverse angiosperms. A similar fossil flora is known from the upper part of the Killik Tongue and from the Niakagon Tongue of the Chandler Formation in the Umiat-Chandler area of northern Alaska (Smiley, 1969a; Fig. 3).

In northeastern Russia, the Grebenka flora from the Krivorechenskaya Formation in the middle reaches of the Anadyr' River (late Albian–Cenomanian to, probably, early Turonian) and middle Ginter taphoflora from the middle part of synonymous formation in the Ugol'naya Bay are most close analogues of the "middle" taphoflora from the Kuk-Kaolak area. The taphoflora from the Yeliseev locality at the Grebenka River, which is most similar in composition, is dated back to the terminal Albian–initial or early Cenomanian (Herman, 1999a, 1999b; Spicer *et al.*, 2002). Species in common for taphofloras under comparison are *Gleichenites* cf. *asiatica, Arctopteris penzhinensis, Taeniopteris* ex gr. *lundgreni, Nilssonia yukonensis, Ginkgo* ex gr. *adantoides, Phoenicopsis* ex gr. *angustifolia, Pityophyllum* ex gr. *nordenskioldii, P.* ex gr. *staratschinii, Araucarites anadyrensis, "Araucarites"* sp. (cone), *Cephalotaxopsis intermedia, Menispermites kryshtofovichi, M.* cf. *reniformis, M.* cf. *septentrionalis, Platanus* cf. *louravetlanica, "Lindera" media, Schefferaephyllum venustum, Grebenkia anadyrensis*, and *Dalembia vachrameevii*. The species known exclu-

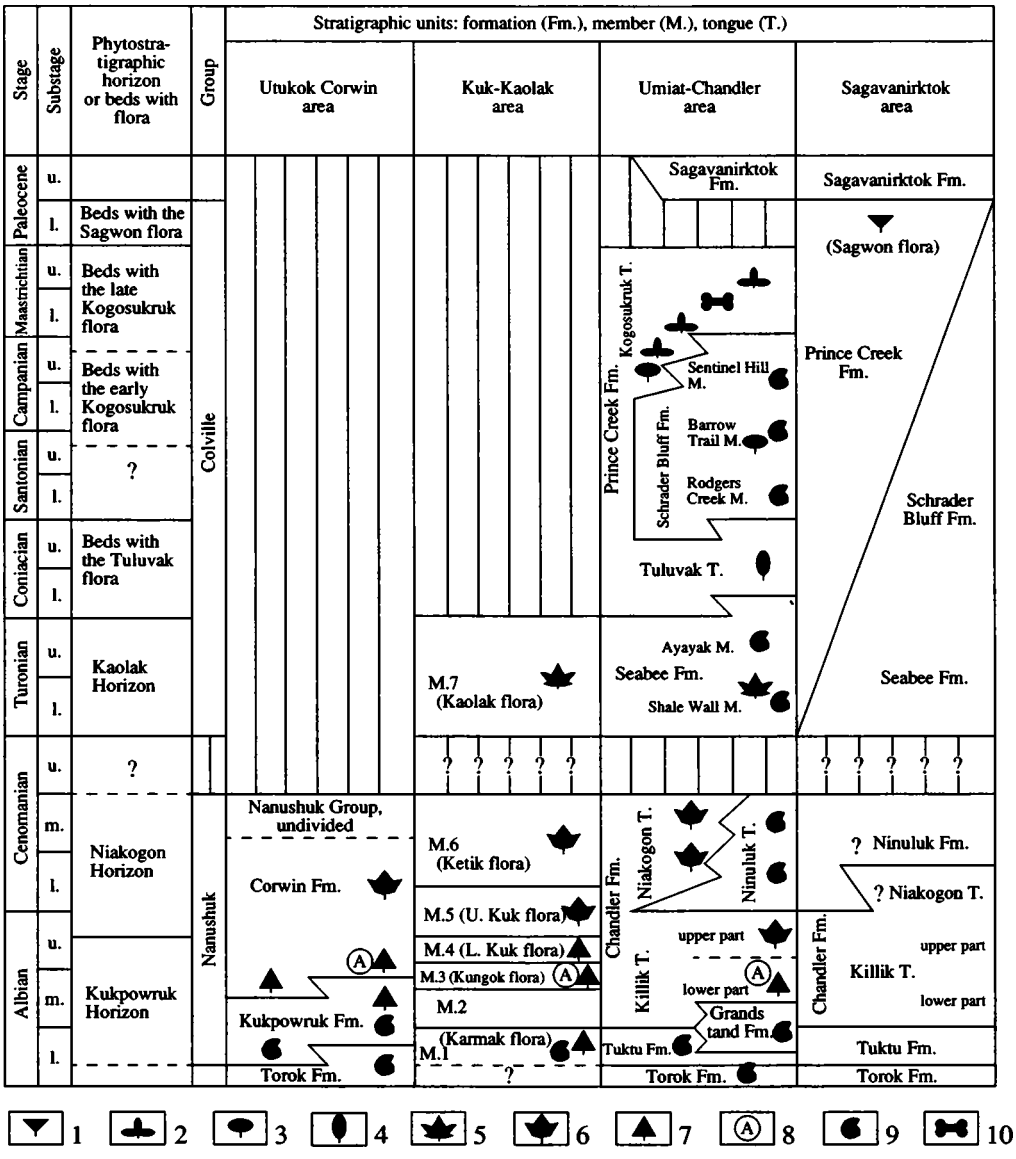

Fig. 3. Correlation of Albian–Paleocene sediments of northern Alaska (phytostratigraphic horizons and beds with flora): (1–7) taphofloras: (1) Sagwon-type, (2) Late Kogosukruk-type, (3) Early Kogosukruk-type, (4) Tuluvak-type, (5) Kaolak-type, (6) Niakogon-type, (7) Kukpowruk-type; (8) first finds of angiosperm plants; (9) finds of marine faunal fossils; (10) finds of dinosaur remains. (l.) lower; (m.) middle; (u.) upper.

sively from these taphofloras are *Gleichenites* cf. *asiatica, Platanus* cf. *louravetlanica, "Lindera" media, Schefferaephyllum venustum, Grebenkia anadyrensis,* and *Dalembia vachrameevii.*

According to stratigraphic position and similarity with the Corwin, Grebenka, and middle Ginter taphofloras, the "middle" taphoflora of the Kuk-Kaolak area appears to be of the late Albian–Cenomanian age (Fig. 3). This inference is inconsistent with conclusions of Smiley who suggested the Cenomanian (Smiley, 1966) or upper Albian age (Smiley, 1972a) to lithological member 5 and attributed member 6 to the Turonian, suspecting therefore a stratigraphic unconformity between them (Table 1). As was mentioned, there are no grounds to suspect unconformity and hiatus in the Kuk-Kaolak section, and changes in taxonomic composition detectable in members 5 and 6 are gradual in my opinion (that is why fossil plants from these lithological units are united into a single taphoflora). The Smiley's assumption (1966, 1972a) that fossil flora from Member 6 ("Ketik flora") is of the Turonian age appears to be misleading because it differs substantially from the Turonian Penzhina flora of northeastern Russia (Herman, 1991, 1999b).

The "upper" taphoflora from the Kuk-Kaolak area is analogous to the Kaolak flora from lithological Member 7 (Smiley (1966). In this taphoflora, I determined 21 species of ferns, conifers, and angiosperms (Table 2).

Fossil wood remains occur in association with leaf impressions. Between two species of not abundant ferns, *Birisia ochotica* is dominant. *Cephalotaxopsis intermedia* is most common taxon of conifers, which are of a low diversity (four species) as well. Much more diverse angiosperms are represented by 14 species of 10 genera. Prevailing among them are genera *"Platanus," Arthollia, "Credneria,"* and *Pseudoprotophyllum* of platanoids and *Trochodendroides*. Representatives of other taxa are rare. It is remarkable that taphoflora under consideration is lacking ginkgoaleans and cycadophytes, while that entire-leaved angiosperms are scarce (one *Leguminosites* impression).

The "upper" taphoflora from the Kuk-Kaolak area is most similar to the Penzhina taphoflora of northwestern Kamchatka, where it is dated back to the Turonian, presumably except for the stage commencement (Herman, 1991, 1999b). The similarity between two floras is evident from the following features: diverse angiosperms with most numerous platanoid forms are dominants; the genus *Trocodendroides*, *T. notabilis* included, appears in both floras; entire-leaved angiosperms are scarce and cycadophytes are absent. In addition to *T. notabilis*, species in common for both taphofloras are *Sequoia tenuifolia, Cephalotaxopsis intermedia, Menispermites* ex gr. *septentrionalis, Platanus* cf. *primaeva*, and *Arthollia* cf. *pacifica*. Genera in common are represented by *Onychiopsis, Birisia, Pityophyllum*, and *Ziziphoides*. Based on this similarity, the "upper" taphoflora from the Kuk-Kaolak area is referred to the Turonian (Fig. 3).

The taphoflora from the Seabee Formation of the Umiat-Chandler area in northern Alaska, where this unit is composed of marine sediments bearing mollusks (Smiley, 1969a), can be considered analogous to the "upper" taphoflora. I determined the following forms among not numerous fossil plants from Smiley's collection: *Podozamites* sp., *Magnoliaephyllum* sp., *Araliaephyllum (?)* sp., *Arthollia* cf. *pacifica* Herman, *Trochodendroides* sp., *Dalembia pergamentii* Herman et E. Lebedev, *Dicotylophyllum* cf. *longipetiolatum* Herman, *Dicotylophyllum* spp., and platanoid leaf fragments. The prevalent angiosperm remains occurring in association with *Arthollia* cf. *pacifica* (platanoid) and *Trochodendroides* forms clearly suggest that the "upper" taphoflora is similar to flora of the Seabee Formation.

The inferred Turonian age of the "upper" taphoflora is substantially older than ranges previously assumed by Smiley: the Senonian to Coniacian presumably (Smiley, 1966) or late Campanian–early Maastrichtian (Smiley, 1972a; Table 1). In northern Alaska, the late Campanian-Maastrichtian age is defined for flora of the Kogosukruk Tongue, which differs form the "upper" taphoflora because of a lower taxonomic diversity (approximately 10 species with sole *Parataxodium wigginsii* being abundant) and dissimilar composition (none species in common) (Spicer and Parrish, 1987, 1990). Although the Coniacian flora from the Tuluvak Tongues of northern Alaska contains some species occurring in the "upper" taphoflora (*Pseudoprotophyllum boreale, Arthollia* ex gr. *pacifica, "Credneria" grewiopsoides, Trochodendroides* ex gr. *arctica*), it is significantly different in general composition and bears a substantial proportion of entire-leaved angiosperms (Herman and Spicer, 1997). The "upper" taphoflora also differs much from the Kuk-Kaolak and post-Turonian floras of northeastern Russia (Herman, 1991, 1999b).

Based on his dating of the Kaolak flora, Smiley (1972a) assumed a substantial hiatus spanning the Santonian and, partly, Campanian intervals between lithological members 6 and 7 in the Kuk-Kaolak area. In my opinion, differences between the "middle" (from members 5 and 6) and "upper" (from Member 7) taphofloras are of the same rank as between approximately coeval Grebenka and Penzhina floras of Northeastern Russia, where they replace each other in the more or less continuous section (Herman, 1991, 1999b). Consequently, one can assume that floras of the Kuk-Kaolak area developed continuously through the Cenomanian–Turonian interval. Although paleobotanic data do not indicate a stratigraphic inconformity between lithological members 6 and 7, a small hiatus (beyond the resolution ability of paleobotanic method) cannot be ruled out in this part of the section. It spans probably the upper Cenomanian and corresponds to the regional unconformity established in the Umiat-Chandler area between the Ninuluk–Niakogon beds and overlying Seabee Formation (Detterman *et al.*, 1963; Smiley, 1966; Fig. 3).

CONCLUSION

The restudy of the Cretaceous flora from the Kuk-Kaolak area allows the following conclusion (Fig. 3).

(1) Three taphofloras are defined in the area. The "lower" one from lithological members 1–4 includes the Karmak and Kungok floras and the lower Kuk subassemblage of Smiley (1966). The "middle" taphoflora characteristic of lithological members 5 and 6 consists of the upper Kuk subassemblage and the Ketik flora of Smiley (1966). The "upper" taphoflora form lithological Member 7 corresponds to the Kaolak flora of Smiley (1966).

(2) In two cases, the taphoflora ages have been substantially changed. The "lower" taphoflora is dated back to the middle-late Albian, the second half of the last substage excluded. Smiley (1972a) who suggested the middle Albian–Cenomanian age for this flora was of a similar opinion in his earlier work (Smiley (1966). The "middle" taphoflora attributed to the late Albian–Cenomanian was considered by Smiley (1966, 1972a) as ranging from the late Albian through Turonian. The "upper" taphoflora is referred to the Turonian; previously it was dated back to the Coniacian (Smiley, 1966)

or even late Campanian–early Maastrichtian (Smiley, 1972a).

(3) The taxonomic composition of studied taphofloras and their ages suggest that Cretaceous floras of the Kuk-Kaolak area developed gradually and continuously. There is no paleobotanic evidence to assume two stratigraphic unconformities between lithological members 5, 6, and 7, as it was thought previously (Smiley, 1972a). Other sections of the Kuk-Kaolak area are also lacking signs of stratigraphic hiatuses. Nevertheless, it cannot be ruled out that members 6 and 7 are separated by an insignificant hiatus undetectable by the paleobotanic method. This hiatus may span the upper Cenomanian and corresponds to the regional unconformity established in the Umiat-Chandler area.

(4) The Cretaceous sedimentation cycle in the Kuk-Kaolak area terminated in the Turonian, not in Maastrichtian as was previously assumed (Smiley, 1972a).

(5) The Tuktu, Chandler, and marine Seabee formations from the easterly Umiat-Chandler and Sagavanirktok areas of northern Alaska can be considered as stratigraphic analogues of lithologic members 1, 2–6, and 7, respectively. The Kuk-Kaolak area is lacking analogues of younger subdivisions known in easterly areas (Prince Creek, Schrader Bluff, and Sagavanirktok formations). Lithological members 1–6 can be correlated with the Kukpowruk and Corwin formations from the westerly Utokok-Corwin area, where younger stratigraphic analogues of Member 7 are unknown.

(6) During the Albian–Paleogene time, geological settings changed gradually and irreversibly so that marine sedimentation and sedimentation in general terminated later in easterly areas of the subregion than in western areas of northern Alaska (Fig. 3).

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Stratigraphy, Geological Events, and a New Model of Late Cretaceous–Early Paleogene Rifting in Central Kamchatka

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Abstract—The stratigraphic scheme for the Upper Cretaceous and Lower Paleogene of central Kamchatka is proposed based on the substantiated age of the Baraba Formation (late Campanian). Similar formations and sequences of the Sredinnyi (Median) and Vostochnyi (Eastern) ranges of the peninsula are correlated. The formation analysis and stratigraphic data are used to compile several paleogeographic schemes. The combined stratigraphic, petrologic, geochemical, and tectonic data constrain a new model of rifting in Kamchatka Peninsula during the considered pre-island-arc period.

Key words: Cretaceous, Paleogene, formation, rift, paleogeography, correlation, volcanism.

INTRODUCTION

In the article, we consider some problems concerning the Late Cretaceous geological history of the southern Sredinnyi Range in central Kamchatka and, for comparison, of several areas in the Eastern Kamchatka Ranges (Fig. 1). Main attention is paid to examination of an obvious stratigraphic marker suitable for further interpretations, even though they are speculative to some extent. In the southern Sredinnyi Range, the Baraba Formation (Baraba Conglomerate) with the diverse floral assemblage appears to be such a marker. Its age and relationship with overlying sediments are debatable so far. Based on the comprehensive study of floral remains from the Baraba Formation stratotype, Chelebaeva concluded that it is the late Campanian in age.[1]

In this work, regional geological events of the transitional Cretaceous–Paleogene period are discussed with due account for tectonic deformations in other structures of Kamchatka and some adjacent areas in order to understand paleogeography through the Campanian–Maastrichtian to Cenozoic time.

UPPER CRETACEOUS STRATIGRAPHY IN THE SOUTHERN SREDINNYI RANGE

Upper Cretaceous deposits are widespread to the north, west, and east of the metamorphic massif located in the southern Sredinnyi Range of Kamchatka. Beginning from the 1960s (*Geology of the USSR...*, 1964), the Upper Cretaceous sequence is divided here into the Baraba, Khozgon, Irunei, and Kirganik formations composed of terrigenous, volcanogenic-terrigenous, and purely volcanic rocks.

Baraba Formation

The Baraba Formation is developed along the northern periphery of the metamorphic massif: around Mt. Baraba and in the watershed area between the Oblukovina River and left tributaries of the Icha River, the Khimka and Khim creeks. Further, outcrops of the formation are observable northwestward along the western slopes of the watershed between the Khimka Creek and the Pozdnii Creek, the next easterly tributary of the Icha River. The formation is usually underlain by quartz–chlorite–sericite crystalline schists of the Khimka Formation or locally by phyllites and metamorphosed sandstones of the older Kheivan Formation. Metamorphic events, which affected these formations, are of unknown age.

Sediments of the Baraba Formation fill in the differentiated relief of the metamorphic basement. The erosion incisions are a few meters to several hundreds meters deep, forming sometimes the small erosion depressions. Thus, different strata of the formation rest upon different layers of underlying sediments.

Being most deformed, a quite complete section of the Baraba Formation is exposed in southern and southwestern slopes of Mt. Baraba along the right bank of the Oblukovina River. We consider this section as the formation stratotype. In the left bank of the Oblukovina River near the Kapitanskaya River mouth, there are exposed conglomerates, which belong, in our opinion, to the Baraba Formation. The attitude of conglomerate beds resting with a sharp unconformity upon the metamorphic complex (Fig. 2) is almost horizontal (dip angles of 5°–10°, less commonly to 15°). The sequence

[1] All relevant problems are thoroughly discussed in the monograph prepared for publication.

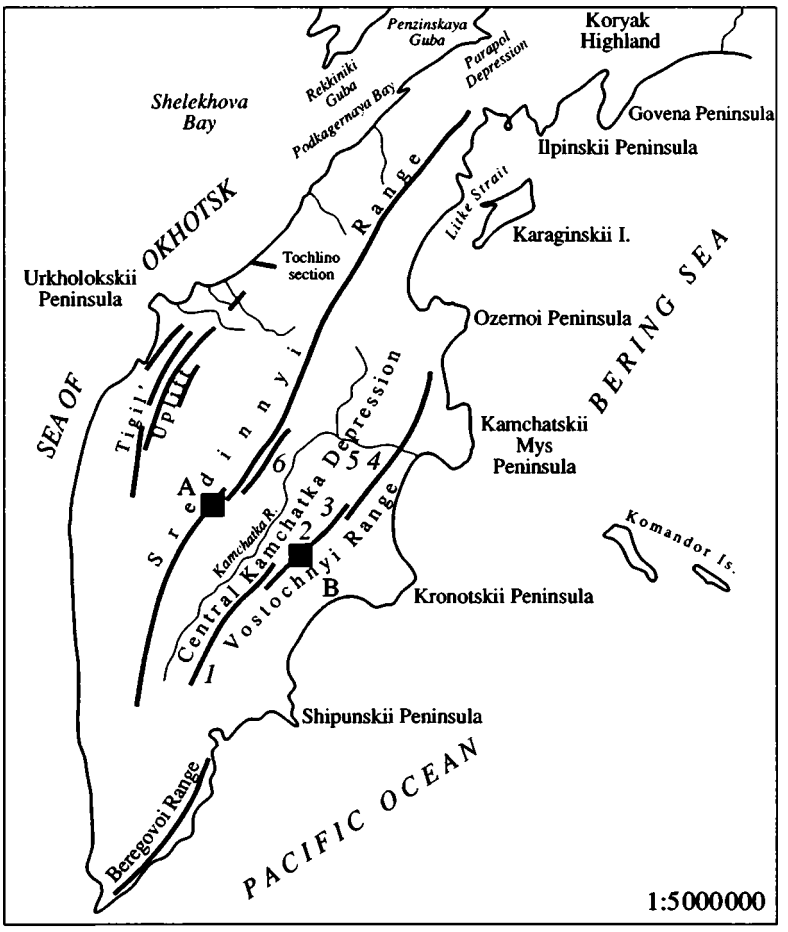

Fig. 1. The geographic scheme of the Kamchatka Peninsula: (A) study area in the Sredinnyi Range; (B) study area in the Valaginskii Ridge; (*1*) Ganal'skii, (*2*) Valaginskii, (*3*) Tumrok, and (*4*) Kumroch ridges; (*5*) Klyuchevskaya group of volcanoes; (*6*) Kozyrevskii Ridge.

consists of four the volcanogenic, conglomerate, conglomeratic breccia, and sandy–silty members. The latter is composed of sandstones, siltstones, and shales. Presented below is characterization of the section (from the base upward).

Volcanogenic member. The member base is exposed along the right tributary of the Kapitanskaya River near its inflow in the Oblukovina River. The intricately deformed schists of the Kheivan Formation are overlain here with a sharp unconformity by compact dark gray aphyric to oligophyric pyroxene–plagioclase basalts. The two lava flows 5 to 10 m thick are separated by tuff interlayer. Higher, the section is largely composed of basaltic, psephitic, and psammitic tuffs and tuffstones (thickness of 20–30 m) grading upward into andesitic tuffs, tuffites, and tuffaceous siltstones (30–35 m thick). The upper part of the member encloses numerous andesite-dacitic and dacitic pyroclastic flows composed of rounded andesite and dacite lava fragments cemented by volcaniclastic matrix.

Remnants of similar dacitic pyroclastic flows are also preserved in ancient incisions in the western spurs of Mt. Baraba. Higher in this section, they grade into the conglomeratic member composed of gravelstones, quartz-feldspar sandstones usually with the volcanomictic admixture, and tuffites grading upward into true conglomerates. Northward and northwestward, the volcanogenic member pinches out. Its maximal thickness is 100 to 120 m.

Conglomerate member overlies conformably the volcanogenic member on the left side of the Oblukovina River and rests with erosional surface on metamorphic rocks on its right side of the Mt. Baraba piedmont (Fig. 2). The member is largely composed of volcanic conglomerates with interlayers of gravely, sandy, and silty tuffs and of mixed andesitic to andesite-dacitic rocks. Medium-, coarse-, and boulder-sized pebbles are well rounded; subrounded and subangular rock fragments are infrequent. Pebbles and boulders are mostly composed of dacite or andesite, which are light-colored, with rare phenocrysts of plagioclase (oligoclase–andesine Nos. 25–30) and color minerals (pyroxene and hornblende) replaced by the epidote–chlorite aggregate. Pebbles of acidic tuffs, granites, granodior-

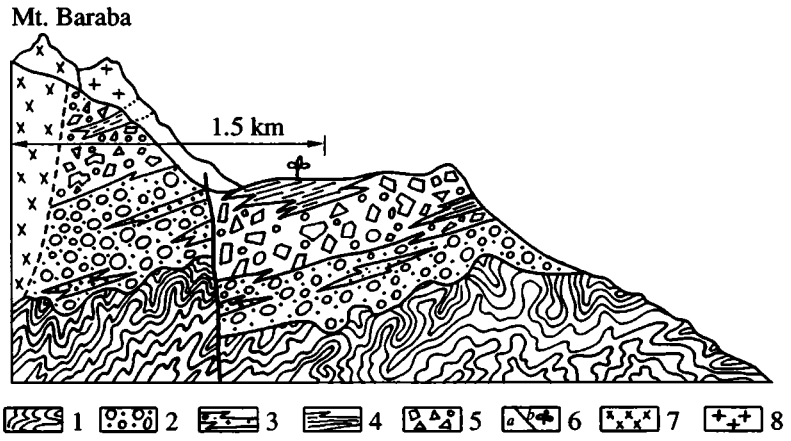

Fig. 2. Geological section of the Mt. Baraba northern crests: (1) metamorphic rocks; (2) conglomerates; (3) sandstone and gravelstone lenses; (4) siltstone lenses; (5) conglobreccia (third member); (6) faults (*a*) and fossil flora finds (*b*); (7) contact hornfels zone; (8) granitoid intrusion.

ites (porphyric varieties), dolerites, tuffstones, reddish brown tuffaceous siliciliths, phyllites and metamorphosed basalts are subordinate. Crystalline schists of the basement are rare among pebbles. Sandy cement of conglomerates frequently shows admixture of tuffaceous material; shapes of glass clasts are characteristic of volcanic ash. The member thickness is approximately 120–150 m and decreases to zero northwestward.

Conglomeratic-breccia member is composed of subangular and angular fragments and boulders (up to 10–15 m across) of metamorphic rocks. The poorly sorted fragments form sometimes boulder breccia, variably metamorphosed similar to rocks of the Khimka and Kheivan formations underlying the Baraba Formation. Rocks observable in coarse clasts are slightly recrystallized sandstones, siltstones, and tuffaceous siliciliths, as well as graphite–mica–quartzose to albite–mica–chlorite schists and gneisses. Less common are pebbles of intermediate to acidic volcanics and granite-porphyry.

Relationships between the second and third members are observable in many areas. The boundary between conglomerate beds dominated by pebbles of volcanic or metamorphic rocks is always sharp without a kind of transitional lithology. On the other hand, proportion of metamorphic rock fragments is increasing toward the second member top, while the basal part of the third member contains abundant fragments of volcanic rocks. In the stratotype section, the maximal thickness of conglomeratic-breccia member is up to 550 m.

Sandy–silty member is defined arbitrarily because it grades laterally and vertically into the conglomeratic-breccia member. Main rocks of the member are alternating gravelstones, sandstones, and siltstones with subordinate small-pebbled conglomerate interbeds. Sandstones and siltstones are polymictic (quartz, plagioclase, metamorphic rocks), with a minor admixture of volcanogenic material. In the uppermost part of the section, sandstones and siltstones show a vague graded bedding. East of Mt. Baraba, the member is intruded by Miocene granodiorites. A zone of black amorphous contact hornfels after uppermost sediments is quite wide. The member is of a variable thickness, of 70–100 m at maximum. In stratotype section, the Baraba Formation is approximately 800–900 m in total (Fig. 2).

Abundant plant detritus large- to small-sized is dispersed throughout the entire formation section. Most significant accumulations of leaf flora and carpological remains are found at two levels in rocky outcrops of the basal part of the sandy–silty member on the southwestern slope of Mt. Baraba. The lower layer contains mostly carpoliths, while leaf impressions are frequent in the upper one. We consider plant fossils from these layers as the stratotype flora of the Baraba Formation. In total, this flora includes 41 species of 35 genera: *Nilssonia yukonensis* Hollick, *Williamsonia kamtschatica* Cheleb., *Araucarites* sp., *Torreya* sp., *Metasequoia disticha* (Heer) Miki, *Magnolia ochotica* Budants., *Menispermites kryshtofovichii* Vakhrameev, *Cocculus* sp., *Tetracentron barabensis* Cheleb., *Platanus basicordata* Budants., *P. primaeva* Lesq., *Arthollia evenensis* Cheleb., *A. insignis* Herman, *Paraprotophyllum ignatianum* (Krysht. et Baik.) Herman, *Macginitiea kamtschatica* Cheleb., *Ulmus schanceri* Cheleb., *Fagus aspera* Cheleb., *F. echinata* Cheleb., *Vlamenneria lucida* Cheleb., *Corylus beringiana* (Krysht.) Golovn., *Juglans* sp., *Grebenkia magna* Cheleb., *Ternstroemites shapiroi* Cheleb., *Acer milkovense* Cheleb., *A. priscum* Cheleb., *A. rubifolium* Cheleb., *Beringiaphyllum ancestralis* Cheleb., *Rhamnus indigenus* Cheleb., *Cissites* cf. *beljaevii* Herman, *Dicotylophyllum* sp. 1–2, *Nyssidium arcticum* (Heer) Iljinsk., *Carpolithes barabensis* Cheleb., and others. Most of them represent form genera, probably extinct. Ten forms are referred to recent angiosperm genera and many of them arbitrarily

because they are represented by species lacking among recent plants.

Leaf impressions mostly belong to denticulate-leaved species; some of them characterize difficultly identifiable entire-leaved forms. Impressions of *Arthollia, Vlamenneria*, and *Beringiaphyllum* are dominant. One interbed contains abundant *Nilssonia*. Leaf remnants are medium to very large in size (up to 25–40 cm long), particularly large are leafs of the *Magnolia, Arthollia, Vlamenneria*, and *Grebenkia* genera. The flora is mesophilic with thermophilic elements represented by *Nilssonia, Williamsonia, Magnolia, Macginitiea*, and *Ternstroemites*.

The comparison of the Baraba floral assemblage with floras from Upper Cretaceous phytostratigraphic units of northeastern Russia (Herman and Lebedev, 1991; Herman, 1993, 1994; Filippova and Abramova, 1993; Golovneva, 1994) elucidates its stratigraphic position.

In its general appearance and presence of different taxonomic groups, the Baraba flora is comparable with the mesophilic Turonian–Coniacian and Maastrichtian floras. It includes forms close to or identical with species from the Penzhina, Kaivayam, and Gornaya Rechka phytostratigraphic horizons. These are *Arthollia insignis, A. evenensis, A. pacifica, Paraprotophyllum ignatianum, Platanus primaeva, Menispermites kryshtofovichii, Acer rubifolium*, and others. *Grebenkia* represents likely a relict of the early Turonian (Cenomanian, according to recent data) floras, although the studied flora cannot be regarded as analogue of any of them. In composition, the Baraba flora is transitional and includes species that occur only either in the Penzhina and Kaivayam assemblages (*Artholia insignis, Paraprotophyllum ignatianum*, and others) or in the Gornaya Rechka flora (*Acer rubifolium*). Only the Barykov phytohorizon corresponding to the Santonian and lower (or lower–middle) Campanian is known between above assemblages of mesophilic floras.

The small- to narrow-leaved Barykov flora with a significant proportion of entire-leaved forms is indicative of the climatic optimum with a notable deficiency of atmospheric moisture. The Baraba and Barykov floras are different in many aspects, although they have some genera in common (*Nilssonia, Magnolia, Acer*). It can logically be assumed that the former is the post-optimum (post-Barykov) time that was close in climatic parameters to the Gornaya Rechka epoch, although flora of the latter is richer in relict forms and thermophilic elements. Taking into consideration the known data on the early Maastrichtian cooling, we think that the Baraba flora is most likely the late Campanian in age.

Khozgon Formation

The Khozgon Formation is exposed in the Khimka River headstream and upper reaches of the Khim River north and northeast of Mt. Baraba. The largest exposures of the formation are observable on the left side of the Oblukovina River downstream of the Bogdanovskaya River mouth. Unfortunately, it is impossible to compile its continuous section because of rare exposures and strong deformations. There are small westward overturned folds up to 30 m wide. The sequence is disturbed by numerous fractures from normal faults to thrusts that were formed in the Paleogene and Late Cenozoic. Because of this situation, only general characteristic of the Khozgon Formation section is possible.

The formation is composed mainly of black clayey siltstones, shales, and fine-grained sandstones composed of quartz and oligoclase grains and clasts of metamorphic and siliceous rocks. Local hieroglyphs and graded stratification are observable in sandstone beds. The above rocks are intercalated with lenses and laterally variable interbeds of volcaniclastic gravelstones with fragments of andesite and basalts, greenish gray siliceous shales, black to gray cherts with radiolarians and inoceram prismatic layers. The formation encloses also polymictic conglomerate beds similar in composition to those in the Baraba Formation (Zinkevich *et al.*, 1994). Several large lenticular flows of tholeiitic pillow lavas are also present.

Stratigraphic relationships of the Khozgon Formation with underlying sediments are unknown. In some areas, the formation is thrust over either metamorphic rocks (Pozdnii Creek basin, low-angle thrust fault) or the Baraba Formation (Khimka River basin, steep thrust fault). In the hanging wall of the latter fault, the formation is composed of polymictic fine-grained sandstones and siltstones. Noteworthy is that the upper part of the visible section of the Baraba Formation in the lying wall is represented by identical rocks. It should be noted that transition between the Baraba and overlying Khozgon formations is gradual in this area (I.A. Sidorchuk, private communication). Taking into consideration the data by Khanchuk (1982) on the transgressive attitude of the Khozgon Formation on metamorphic rocks in southerly areas of the Sredinnyi Range western slope, basal thrust sheets can represent the detached small blocks of both the metamorphic basement and the Baraba Formation. Thrusting of this kind could be related to gravity processes and different rock rheology. The total formation thickness is unknown. Some researchers assume that formation can be as thick as 2000 m (*Geology of the USSR*, 1964). Based on radiolarians, the formation age is determined as either middle or late Campanian (Shapiro *et al.*, 1986).

Irunei Formation

The Irunei Formation is the most widespread stratigraphic unit in the area under consideration. It is of diverse facies composition, although clayey-siliceous and clayey-tuffaceous rock varieties are dominant. In the Khim and Oblukovina river basins, the formation is

largely composed of green-gray silty and silty-clayey tuffs grading into siliceous tuffites and greenish gray tuffaceous siliciliths; locally, members and lenses of cherry–red siliceous argillites with jasper intercalations occur. In the Pravaya Andrianovka River basin, the upper part of the formation composed of the above-mentioned rocks encloses psammitic and, sometimes, coarser tuffs of the basaltic composition, which are intercalated with rare thin lava flows. Siliceous-clayey fine-grained rocks contain abundant fragments of inoceram prismatic layers and less common large poorly preserved impressions of their valves.

In the Bogdanovskaya River basin, the presumably upper part of the formation composed of tuffaceous rocks encloses a member of black siltstones and shales with a significant arkose admixture. By composition, siltstones are similar to those from the Khozgon Formation. Because of tectonic deformations, it is impossible to compile the continuous section of the formation. Nevertheless, Zinkevich *et al.* (1994) managed to subdivide the formation, probably arbitrarily, into the clayey-siliceous, siliceous-tuffaceous, and tuffaceous sequences with tectonic contacts between all of them. In different parts of the area, these rocks form separate tectonic sheets. The total thickness of the formation is estimated to be 700–900 m. Based on radiolarians, V.S. Vishnevskaya determines its age as the late Campanian–Maastrichtian (Shapiro *et al.*, 1986). L.G. Bragina estimated a similar stratigraphic range (Zinkevich *et al.*, 1994). These age determinations indicate facies connections between the Khozgon and Irunei formations.

Kirganik Formation

Outcrops of the Kirganik Formation are known east of Mt. Baraba in upper reaches of the Bogdanovskaya River, along the eastern fringing areas of metamorphic rocks exposed along the Kirganik, Zhupanka, and Andrianovka rivers, and farther southward up to the Ozernaya Kamchatka River. The formation is mainly composed of agglomerate, psephitic, and, less commonly, psammitic tuffs. In typical section, prevailing clastic material of subalkaline to alkaline basalts and pycrite is mixed with rudaceous tuffs of the same composition. At the watershed between the Pravaya Andrianovka and Zhupanka rivers, volcanics of the Kirganik Formation are gradually replaced upward the section by the sedimentary terrigenous–siliceous sequence. Normal stratigraphic contacts with underlying Irunei Formation have not been found northeast of Mt. Baraba. In some blocks, there are transitional sections, where proportions of rudaceous volcanic rocks and basalts quickly increase upward the Irunei Formation section. It is likely therefore that the Irunei and Kirganik formations are interrelated via transitional beds. Rotman (1961) arrived at the same conclusion. Marchenko (1959) considered the Kirganik and Irunei formations as facies analogues. We admit the partial lateral replacement of the uppermost part of the Irunei Formation by the lower part of the Kirganik Formation, although the latter, as a whole, is younger. Flerov, who studied for a long time the structure and lithology of the Kirganik Formation, subdivides it into the lower volcanogenic, middle sedimentary, and upper volcanogenic parts (Flerov and Koloskov, 1976). According to our data, sedimentary rocks are characteristic of the uppermost part of the section (probably, the fourth member).

Benthic foraminifers of wide stratigraphic range suggest the Maastrichtian age of the Kirganik Formation (Pozdeev and Petrina, 1984). According to Bragina, radiolarians from the upper terrigenous–siliceous member are Maastrichtian–Danian in age (Zinkevich *et al.*, 1994). K–Ar dates obtained for gabbro–pyroxenites, the comagmatic rocks with respect to Kirganik volcanics (Flerov and Koloskov, 1976), range within the Paleogene time span. These dates of 52 ± 3, 55 ± 3, and 53 ± 1 Ma (biotite fractions) and 56 ± 4 (whole-rock sample) were obtained at the Laboratory of Isotopic Age of the Institute of Geology of Ore Deposits, Petrography, Mineralogy, and Geochemistry, Russian Academy of Sciences. The age can be slightly overestimated because magmatic biotite is associated in the rocks with abundant later metasomatic biotite. The thickness of the Kirganik Formation is variable, ranging from 600 to 1200 m; some volcanic members probably pinch out.

Because of tectonic relationships between formations mentioned above and limited paleontological materials (Baraba Formation is an exception), it is difficult to figure out their composite succession based on indirect evidence from transitional facies analysis and extremely rare stratigraphic contacts. Nevertheless, 100–120 km east–southeast of the area under consideration, there is a section of similar lithology and age in the well-exposed Valaginskii Ridge, where constituting subdivisions overly each other with conformable stratigraphic contacts. Therefore, it seems reasonable to correlate sections of the Sredinnyi Range and Valaginskii Ridge.

CORRELATION BETWEEN SECTIONS OF THE SREDINNYI RANGE AND VALAGINSKII RIDGE

Stratigraphy of the Valaginskii Ridge was studied by many researchers in the period from 1959 to 1993 (V.V. Tikhonov, M.I. Goryachev, A.E. Shantser, A.G. Tsikunov, I.V. Florenskii, V.S. Petrov, M.E. Boyarinova, B.I. Slyadnev, V.A. Seliverstov, M.K. Bakhteev, V.P. Zinkevich, and others). Cretaceous–Paleogene sediments of the ridge are best exposed in the Valagina, Kitil'gina, Levaya, and Vakhvina river basins.

The Golubov and Kitil'gina formations defined in the area are united into the Valagina Group (corresponding to the Khapitsa Formation in the northern Vostochnyi Range and Kumroch Ridge). The group is

overlain conformably and with a gradual transition by the Tal'nikovaya terrigenous formation. It overlies, likely with a gradual transition as well, the terrigenous Vakhvina Formation of a debatable stratigraphic status. Presented below is a brief characterization of composition and age of the indicated units.

Vakhvina Formation. The unit was always of an ambiguous stratigraphic status. In opinion of some researchers, the unit underlies the Valagina Group being of the same rank (*Resolutions of the Second Interdepartmental...*, 1982; Zinkevich *et al.*, 1992), while others place it below the Kitil'gina Formation under the name Vaskuchev Formation (Petrov *et al.*, 1966). Accordning to next viewpoint, two terrigenous sequences underlying and overlying the Valagina Group are erroneously united into the Vakhvina Group (Seliverstov and Grechin, 1979). Finally, M.E. Boyarinova believes that the Vakhvina and Tal'nikovaya formations constitute a single terrigeonous complex overlying the Valagina Group. Previously, we shared the last viewpoint as well (Bakhteev *et al.*, 1994). Nevertheless, Slyadnev (1999) clearly showed that in the southern Valaginskii Ridge (Vakhvina River, Lake Sevo) the Vakhvina Formation underlies conformably the Kitil'gina Formation with a visible transition from subarkose sandstones to tuffs and tuffaceous–siliceous rocks (this author considers the Golubov and Kitil'gina formations as a single stratigraphic unit). Slyadnev also described a gradual transition of the Vakhvina Formation into the lower part of the Valagina Group and presented data on the different mineral and chemical compositions of the Tal'nikovaya and Vakhvina formations. We should agree that his standpoint seems to be most substantiated now.

Unfortunately, the age of the terrigenous Vakhvina Formation is unknown yet. Taking into consideration its gradation into the Golubov Formation of the Campanian–Maastrichtian in age (see below), we suggest the Campanian age for the Vakhvina deposits. The thickness is up to 2000 m.

Golubov Formation. In the Valagina River section, the formation corresponds to the basal section part, where it is composed of gray to greenish gray horizontally laminated tuffaceous–siliceous and clayey–siliceous rocks with interbeds and members of gray, black, greenish-gray, and brown cherts. Basaltic tuffs, tuffstones, and siliceous tuffites are subordinate in abundance. The proportion of basaltic tuffs increases in the upper part of the formation. Rocks show variable propilitization and carbonatization.

According to N.Yu. Bragin, the radiolarian assemblage from the middle part of the Golubov formation is the late Campanian–Maastrichtian in age. In opinion of D.I. Vitukhin, some radiolarian taxa are characteristic of the earliest Paleocene as well (Bakhteev *et al.*, 1994). Accordingly, the Golubov Formation is arbitrarily attributed to the late Campanian–early Paleocene. Its thickness exceeds 600 m.

Kitil'gina Formation. The Golubov Formation is overlain with a gradual transition by the Kitil'gina Formation of a two-member structure. Its lower, substantially pyroclastic part (approximately 70% of the formation section) is composed of greenish gray psammitic, vaguely bedded basaltic agglomerates and tuffs with scarce basaltic flows and interbeds of lava breccias and clastic lava, which are intercalated with thin lenses of greenish chert. The formation upper part shows a lower proportion of pyroclastic material, being mainly composed of alternating basaltic flows and tuff interbeds. Lavas and tuff-agglomerates contain abundant inclusions of gabbro, gabbro–pyroxenites, and dunites. The formation is barren of fossils being up to 1300 m thick.

Tal'nikovaya Formation conformably overlies the Kitil'gina Formation. The transitional member is thick-bedded, dominated by sandstones, gravelstones, and siltstones, while proportion of tuffs and tuffites is significantly decreased. Its other characteristic feature is presence of interbeds composed of poorly sorted sandstones enriched in gravel and finer clasts of basalts, limestones, and quartz. The member is up to 300 m thick. As a whole, the Tal'nikovaya Formation is sandy–clayey in lithology. We divide it into four members, one of which (transitional) is characterized above. The second member (approximately 400 m thick) reveals a cyclic alternation of thick (30–50 m) black silty–clayey beds and gray to bluish gray quartz–feldspar and, less commonly, volcaniclastic sandstone beds 3 to 10 m thick.

The third member (approximately 200 m) is largely composed of two- or three-component flysch of the feldspar–quartz–graywacke composition. Both the graded and parallel cross-bedded structures are typical of the member beds. Rhythmical patterns are frequently interrupted by coarse-grained sandstone interbeds from 1 to 3 m thick. The fourth member (190 m thick) is dominated by black argillites and siltstones with thin intercalations of subarkose sandstones enclosing limestone concretions.

The middle part of the Tal'nikovaya Formation yielded the planktonic foraminifer assemblage indicating the late Paleocene–early Eocene age of host sediments (Bakhteev *et al.*, 1994), although a wider range, the early Paleocene included, cannot be ruled out. The total thickness of the formation is 1000–1100 m.

The Valagina Group is distinguishable practically all along the Vostochnyi Range (in the Kumroch Ridge northward, its equivalent is the Khapitsa Formation). The group is composed of uniform volcanogenic–siliceous rocks in its lower part and of volcanics in the upper one. Almost everywhere, the group is conformably overlain by terrigenous, usually flyschoid sequence (Tal'nikovaya Formation, Drozdovo Formation). In different areas of the Vostochnyi Range, the group is voluntarily divided into formations and sequences with different, sometimes confusing names.

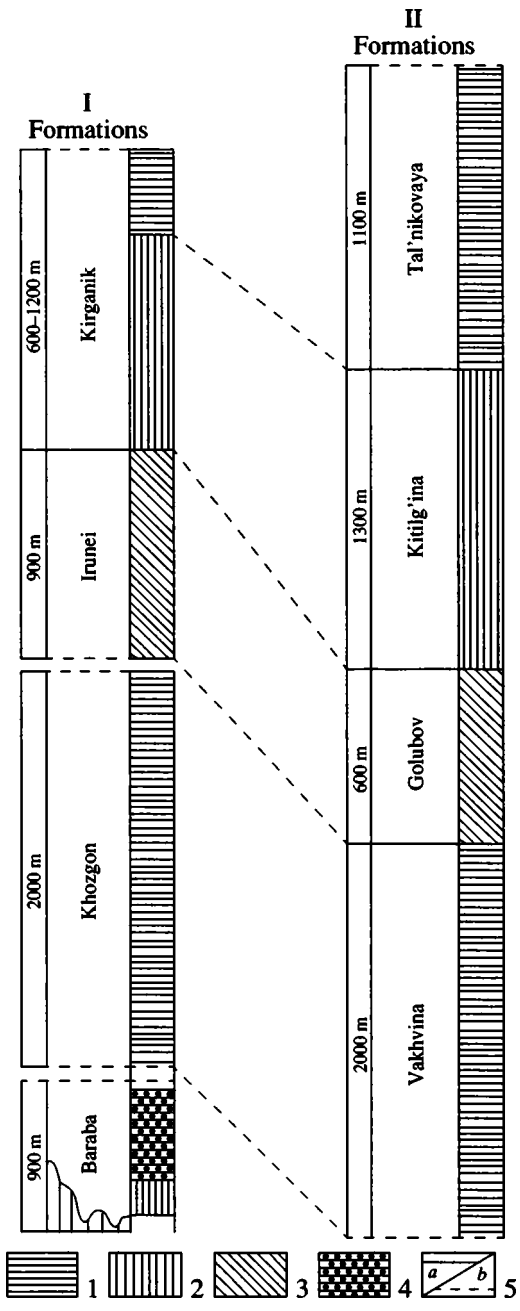

Fig. 3. Succession of formations in Upper Cretaceous–Lower Paleogene sections of the Sredinnyi Range (I) and Valaginskii (II) Ridge: (1) terrigenous, (2) mainly volcanogenic, (3) siliceous–volcanogenic, (4) terrigenous–volcanogenic molasse formations; (5) conformable (*a*) and tectonic or unknown (*b*) contacts.

Despite this ambiguous situation and scarcity of biostratigraphic information, there is a certain similarity in succession of different formations in the Upper Cretaceous–Paleogene sections of the southern Sredinnyi and Vostochnyi ranges, the Valaginskii Ridge included. According to lithology and stratigraphic data, the Irunei Formation of the Sredinnyi Range seem to be an analogue of the Golubov Formation of the Valaginskii Ridge, while the Kirganik Formation corresponds, based on the same criteria, to the Kitil'gina Formation. The Vakhvina Formation can be an age analogue of the Khozgon Formation (Fig. 3).

PALEOGEOGRAPHY

We realize that any paleogeographic interpretations for the Late Cretaceous will be arbitrary because of insufficiently elaborated regional stratigraphy. Nevertheless, we would like to discuss below a model, which seems to be more or less plausible for the region under consideration.

Baraba–Khozgon time. As was mentioned, the Baraba Formation is composed almost entirely of continental facies, which recorded the beginning of late Campanian–early Eocene development cycle of Kamchatka after the middle Campanian orogenic phase and intense erosion (Gladenkov *et al.*, 1997). It is likely that at the commencement of Baraba Formation accumulation there was a belt of subaerial contrasting basalt–dacite volcanism superimposed on the metamorphic basement west and, probably, south of the formation distribution area. The material eroded from this belt formed the second member of the formation, which is almost entirely composed of volcanic conglomerates with rare tuff interbeds. Changes in topography of metamorphic basement were accompanied by changes in erosion areas and direction of the detrital material transport, as it is reflected in sharp lithological changes of the third member largely composed of metamorphic rock clasts. The appearance of siltstones and, even, shales in the upper part of the formation section implies a sharp decrease in erosion intensity and emergence of small water basins with calm hydrodynamics, where environments were favorable for autochthonous burial of leaf remains. The lateral facies transition of the Baraba Formation (especially of its upper layers) into the Khozgon Formation cannot be ruled out. At any rate, we consider the Baraba Formation as the basal unit of the overlying sequence, which recorded the initial sedimentation stage in the forming sea basin (Fig. 4).

In *the Khozgon time* of transgression, the sea flooded southern depressions in the metamorphic basement of the Sredinnyi Range. Sedimentation during this epoch was mostly terrigenous, and detrital material was transported from islands composed of metamorphic rocks and from the Okhotsk land, the fold zone of Mesozoic (pre-late Campanian) and metamorphic rocks, which was peneplanated at that time. The quantity of arkose material in the Khozgon Formation is substantially high. The Khozgon sea floor was uneven, with separate depressions filling in with turbidites.

Tuffs, mixed tuffaceous–terrigenous rocks, and rare lenses of tholeiitic basalts intercalated with predominantly terrigenous section suggest episodic eruptions from small shield-shaped volcanoes at the sea floor.

Rare chert interbeds can be of both the biogenic and volcanic (altered vitroclastic tuffs) origin.

Difficulties in paleogeographic interpretation are related to the Khozgon Formation that is scarcely exposed and ambiguously correlated. If the Lesnaya Formation of the Kamchatka Isthmus (entirely or partly), Vakhvina Formation of the Valaginskii Ridge, black-shale sequence of the northern Ganal'skii Ridge, and several Upper Cretaceous terrigenous sequences observable or drilled in western Kamchatka (Antipov *et al.*, 1998) can be considered as correlative with the Khozgon Formation, one can assume that in the study region there was a marginal sea at that time, probably with isolated islands composed of Mesozoic and metamorphic rocks. In addition to known exposures of metamorphic rocks (Sredinnyi Range, Ganal'skii Ridge, Khavyvenka Uplift), it is possible to outline a land area of metamorphic rocks in the southern part of the Kamchatka Peninsula, where fragments and even blocks of crystalline schists occur in alluvial–proluvial sediments of the Fal'shivaya River flowing from the Mutnovskii Volcano area. This clastic material is probably derived from a southern continuation of the Ganal'skii Ridge metamorphic block now hidden under volcanic products. The suggested relationships between land and sea areas are shown in the paleogeographic scheme (Fig. 4).

The Irunei time. By analogy with the Valaginskii Ridge section, the Irunei Formation is most likely younger than the Khozgon Formation, although the lateral facies transition is admissible for the formation lower beds. The Irunei Formation and sequences of comparable age and facies are laterally more widespread than the Khozgon Formation, although insufficiently for a reliable paleogeographic map. We can only suggest that this was a time of extended transgression and subsidence of most islands. In the final Irunei stage there was likely a slight regression or, at least, a basin shoaling despite numerous depressions in the sea floor. The inference is based on numerous remains of inoceram prismatic layers occurring in the formation. Intact inoceram shells, usually large and thick-walled, are indicative of shallow settings, and their prismatic layers represent coquina lenses subjected to calcitization. In addition, the formation contains psephitc tuffaceous and terrigenous material (even in the uppermost agglomerates) which could hardly be deposited at a significant depth. It should also be noted that siliceous material does not form noticeable layers. It is typical of isolated lenses that accumulated probably in some seafloor depressions separated from land by bottom uplifts, which sheltered depression from terrigenous material influx. Dissimilar to the Khozgon time, the Irunei period was characterized by the intense volcanism largely of basaltic, calc-alkaline to tholeiitic composition. Like before, numerous and small shield volcanoes erupted presumably these lavas.

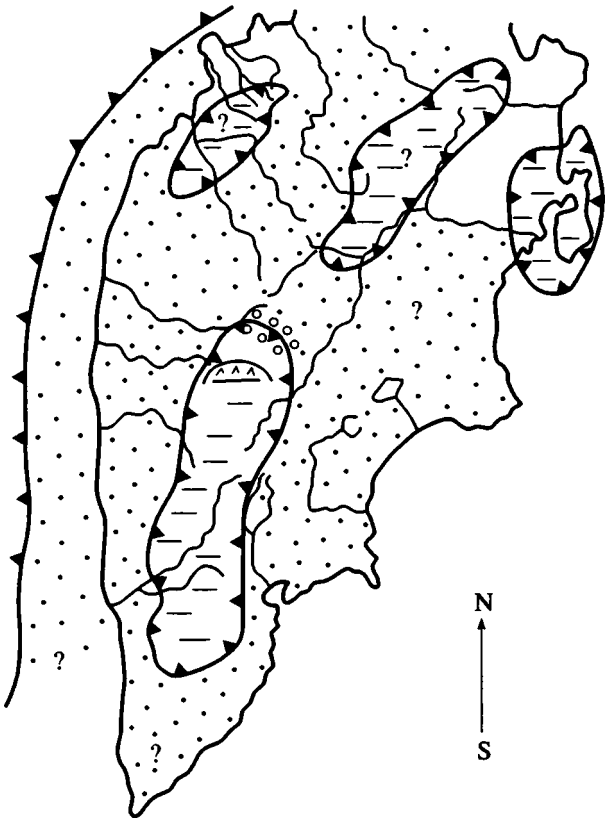

Fig. 4. The paleogeographic scheme of the southern Kamchatka Peninsula for the late Campanian (Baraba–Khozgon time, symbols as in Fig. 6).

The sea was probably spread over a large part of Kamchatka, occupying the Sea of Okhotsk coastal areas, where terrigenous facies prevail. Judging from known exposures of Upper Cretaceous volcanogenic–siliceous formations in the Kronotskii Peninsula (Mys Kamenistyi Formation), Avachinskaya Bay area and in the Kamchatskii Mys Peninsula (Tarkhov Formation?), the sea was connected with the Pacific in the east (Fig. 5).

The Kirganik time. As is evident from stratigraphic position, biostratigraphic and isotopic data, the Kirganik Formation accumulated in the period from the late Maastrichtian to Paleocene, when magmatism was particularly intense.

A long-term study (Flerov and Koloskov, 1976; Seliverstov *et al.*, 1994; Flerov and Seliverstov, 1999; Flerov *et al.*, 2001) showed that characteristic of the Kirganik Formation stratotype area (from upper reaches of the Kirganik River on the north to the Ozernaya Kamchatka River on the south) and of central and western parts of the Sredinnyi Range is the volcanoplutonic K-basaltoid formation. Among volcanics, there were distinguished two rock associations: subalkaline Pl-shoshonite and plagioclase-free alkaline K-basaltic series. Plutonic rocks of the formation are rep-

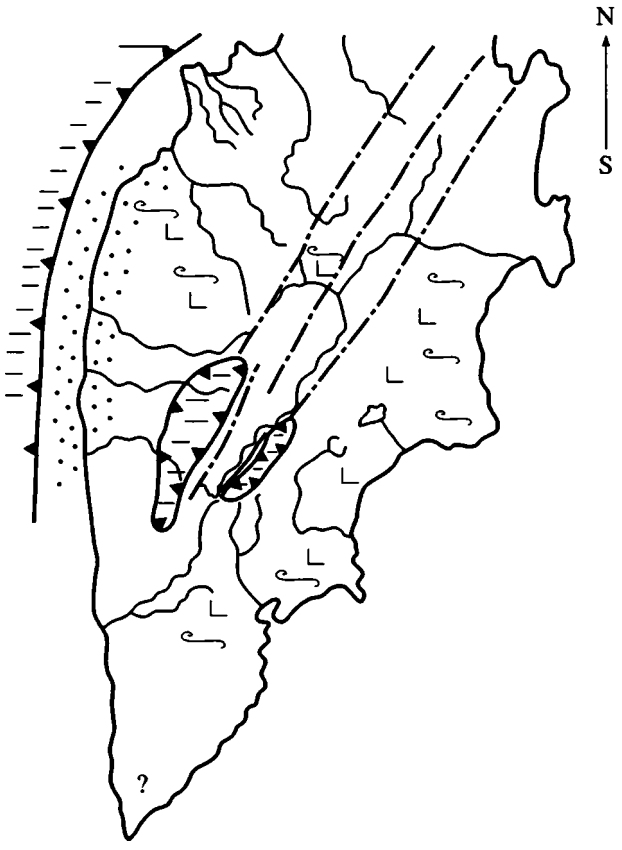

Fig. 5. The paleogeographic scheme of the southern Kamchatka Peninsula for the late Campanian–Maastrichtian (Irunei time). Legend as in Fig. 6.

resented by dunites, pyroxenites, gabbro, syenites, and monzonites. Noteworthy is the fact that the volcanoplutonic formation intrudes both the Irunei Formation and older metamorphic complex.

Another distribution area of similar alkaline rocks is the Vostochnyi Range system of the Valaginskii and Tumrok ridges. Shantser and Erlikh were first to discover here in 1970 the alkaline plagioclase-free volcanics and meimechite-type rocks (Erlikh *et al.*, 1971). Subsequently, this area attracted attention of many researchers (Markovskii and Rotman, 1971; Seliverstov and Tsikunov, 1974; Zinkevich *et al.*, 1993). Selivesrtov (Tomson and Seliverstov, 1992) recognized here the alkaline–mafic–ultramafic formation and cogenetic ring-type layered and intricate dunite–pyroxenite–gabbro–syenite–carbonatite massifs typomorphic of the Kirganik complex. He also described the lamproite-type mineral associations with a single diamond found (Seliverstov *et al.*, 1994).

Based on these materials, Seliverstov and Tomson distinguish the pre-orogenic stage of rifting that can be manifested in the mobile belts and stable structures like the Turan plate. Flerov suggests the rifting stage of that time as well, although within an island-arc system.

It is also important that Sobolev *et al.* (1989) who studied ultramafic rocks of the Valaginskii Ridge considered them as derived from under the continental or subcontinental crust in Kamchatka in the Late Cretaceous and doubted the uniformitarian approach to these and island-arc formations.

As for the lateral distribution of the alkaline mafic–ultramafic formation, it should be noted that lava and pyroclastic eruptions of the Klyuchevskaya volcanoes group (Udina, Zimina, Tolbachik, and others) contain abundant xenoliths of plagioclase-free alkaline rocks, trachybasalts, dunites, and pyroxenites. Similar xenoliths occur also in Pliocene ignimbrites of the Kozyrevskii Ridge (the Sredinnyi Range system). Thus, alkaline mafic and ultramafic rocks were widespread. Lava eruptions were mostly of shallow-water and subaerial types, as it is evident from high proportion of rudaceous pyroclastic material and rare burials of leaf flora in the Kirganik Formation. Fissure eruptions, in addition to those from large volcanic centers, were also common, because this formation is intruded by numerous ring-type radial (relative to volcanic centers) and parallel ultramafic and basaltic dykes in both the Sredinnyi and Vostochnyi ranges.

All these data imply that the Kirganik time corresponds to formation period of a large extension zone (rift) in the sea bottom, which is corresponding now to the central axial zone of the Kamchatka Peninsula. The rifting was associated with intrusion and voluminous eruptions mafic and ultramafic alkaline magma (Fig. 6). Simultaneous accumulation of volcanics and marine sedimentary formations compensated and, sometimes, overcompensated the graben-like subsidence usually characteristic of rift-type structures. It is probable that the central rift formation was accompanied by development of smaller rifts. One of the latter striking northwestward is recognizable in the Malka–Petropavlovsk zone, where its continuation is traceable based on outcrops of ultramafic rocks to the Pacific coast and on ultramafic xenoliths erupted by the Avacha group of volcanoes (Fig. 6).

Simultaneously with emergence of central rift zone, a marginal subaerial volcanic belt started to develop in the extreme western part of Kamchatka in the late Maastrichtian–Paleocene (Shantser and Fedorov, 1999). East and, probably, west of this belt, the coastal paralic coal-bearing formation accumulated (Paleocene). Thus, the western part of the region turned mainly into land probably with small shallow sea bays. At the same time, there were block movements and formation of horsts followed by erosion of Cretaceous and older (metamorphic) rocks that produced (similarly to the Okhotsk land) terrigenous material transported in the eastern direction.

The areas located east of the developing rift accumulated terrigenous sediments (Tal'nikovaya Formation) replaced in easterly deeper parts of the sea by siliceous–carbonate and tuffaceous facies. Basaltic,

mainly tholeiitic or scarcer alkaline volcanism was manifested in some of the areas. The associated carbonate sediments and psephitic tuffs imply shallow depth in this part of the sea. Relationships between the areas and ocean remain unknown so far. It is probable that a chain of volcanic islands or a hypothetical underwater volcanic ridge bounded them on the east during the late Paleocene–early Eocene time.

THE LATE CRETACEOUS–PALEOCENE TECTONIC DEVELOPMENT

In the present-day structure of Kamchatka, Upper Cretaceous sediments are exposed in Late Cenozoic horsts and horst-like uplifts separated by negative structures of various genesis (Shantser, 1987). Many researchers extrapolate the internal fold-nappe structure of the horst to the entire region concealed under Cenozoic sedimentary cover. The thrusting amplitudes are unknown in most cases, and we apply therefore the term fold-and-thrust structure.

For instance, distinct thrust planes and associated zones of rock brecciation and mylonitization may grade into normal stratigraphic contacts along the strike, as it is observable in the well-exposed Valaginskii Ridge, where analogues of the Sredinnyi Range sequences are known. A good example is the thrust plane between the Kitil'gina and Tal'nikovskaya formations in the upper reaches of the Vetlovaya River, where it turns into conformable stratigraphic contact with a transitional member traceable for several kilometers northward. By analogy, we suggest a similar situation for the Mt. Baraba area, where thrust faults are either gravity-related or resulted from irregular lateral compression.

At present, there is a general tendency to explain development of ocean–continent transition zone (Kamchatka in our case) from the viewpoint of plate tectonics (Zinkevich *et al.*, 1993; Shapiro, 1995). We propose the alternative model:

(1) As was mentioned, the volcano-plutonic alkaline mafic–ultramafic formation cannot develop on the oceanic crust.

(2) Ultramafic rocks from the plagioclase-free alkaline ultramafic volcano-plutonic association of the Kirganik Formation intrude both the Irunei Formation and metamorphic rocks of the southern Sredinnyi Range. Consequently, the Irunei and Kirganik formations are spatially connected with the metamorphic complex.

(3) If Slyadnev's data are correct (we are sure they are) and the Vakhvina subarkose formation conformably underlies the Golubov Formation (an analogue of the Irunei Formation in the Sredinnyi Range), it is difficult to imagine accumulation of that formation in open oceanic settings.

We believe that horizontal displacements are not greater than tens of kilometers and the basin width can be estimated straightening folds and "eliminating" thrusts.

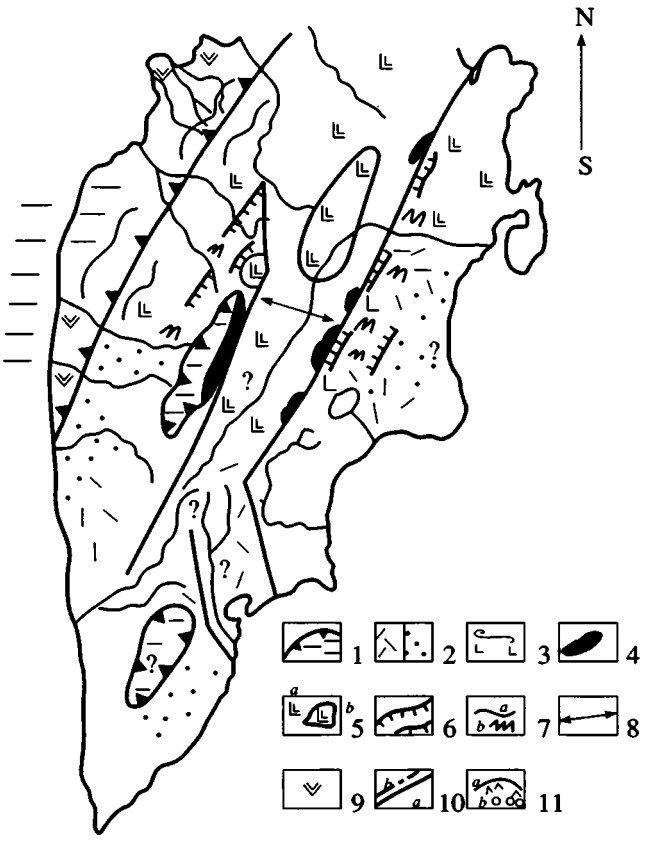

Fig. 6. The paleogeographic scheme for the late Maastrichtian–Paleocene–Eocene (Kirganik time of rifting): (1) land, boundary between land and sea; (2) volcanogenic–sedimentary and terrigenous formations; (3) siliceous–volcanogenic formation; (4) alkaline mafic–ultramafic igneous rocks; (5) distribution areas of alkaline mafic–ultramafic rocks, presumable (*a*) and inferred (*b*) based xenoliths from Holocene volcanoes and Pliocene ignimbrites; (6) thrusts of eastern and western vergency; (7) folding gentle (*a*) and strained (*b*); (8) extension direction; (9) marginal volcanic belt; (10) rift boundaries (*a*) and assumed fracture zones of the Irunei sea floor (*b*); (11) subaerial volcanic eruptions of acidic to intermediate lavas before and partly at the accumulation time of the Baraba Formation (*a*) and molasses (Baraba conglomerates) (*b*).

In the late Campanian–Paleocene time, there was no island arcs similar to present-day Pacific arcs in Kamchatka that experienced rifting instead. A zone of intense extension, which probably involved the present-day Kamchatka depression and partly the Sredinnyi Range, formed in the Irunei sea along a system of deep-seated faults, through which large volumes of mafic–ultramafic alkaline magma intruded the crust (Fig. 6). Precisely, the rifting process determined formation of the fold-and-thrust structure of the region.

The fold-and-thrust structure formed during two stages in submarine conditions. In the first pre-rift stage of synsedimentary folding, the formation of folds was accompanied by small-scale gravity thrusting stimulated by uneven seafloor relief. The second stage of lat-

eral stress resulted in formation of short-amplitude thrusts and inclined folds of western vergence westward of the extension zone and of eastern vergence eastward of the latter.

Kolodyazhnyi (1996), who studied structural geology of the Mt. Baraba area, suggested a subhorizontal westward shear flow that resulted from thrusting of Upper Cretaceous complexes over metamorphic rocks. Shear deformations are also observable at the base and inside the Baraba Formation. We consider them as local deformations that resulted from general compression in the area adjacent to the rift. The rift developed up to the Eocene, which is indirectly evident from presumably Eocene volcanics found in the northern part of the Kamchaka depression (unpublished data of E.V. Daragan and V.A. Uspenskii).

The folded orogenic land, i.e., the true orogen could hardly be formed during a brief period of land rise in the middle Eocene. The rising movements involved the already formed fold system, when rifting ceased in its central part. The associated local deformations were most likely of gravitational origin. These are probably the folds of eastern vergence in the periphery of metamorphic rocks and some thrust faults.

We believed previously that the fold-and-thrust structure of Kamchatka formed during two brief tectonic episodes: in the lower–middle Eocene and late Miocene. Now, we think that the process was slower, lasting millions of years immediately in the sedimentation basin. Folding and thrust deformations increased thickness of sedimentary cover, and the land rise above sea level only marks termination of each of the stages during the long-lasted process. As was mentioned, Seliverstov termed the Campanian–Paleogene development stage of the region as pre-orogenic. It is hardly so. A true orogen has not been formed by the end of this stage, when a temporary uplift appeared was quickly eroded and transgression of a shallow sea began to develop again.

It should be noted that rifting in Kamchatka was reactivated in the Miocene–Pliocene being still in progress. The rift development determines formation of the general peninsula structure (Ermakov, 1987). Thus, the entire geological interval from the Late Campanian to present should probably be considered as the pre-island-arc stage, because Kamchatka is now not yet an island arc of the Pacific type (this issue will be discussed elsewhere).

The proposed model of rifting and related fold-and-thrusts deformations is preliminary of course (like each tectonic model), as the factual material is insufficient. Further studies will probably show how valid the model is. Now, it seems preferable for us at least.

CONCLUSION

In conclusion, it should be emphasized that the most important aspect of our work is the Baraba Formation dating back to the late Campanian based on the studied paleoflora. Indirect evidences suggest a similar succession of lithologic formations in the Sredinnyi Range and Valaginskii Ridge.

All other inferences are preliminary and hypothetical with respect to both the tectonics and paleogeography. Based on factual materials (though insufficient), intuition, and long-term study of Cretaceous to Holocene sections in the region, we proposed an alternative model of region development during the late Campanian–Eocene, in which the rifting of continental or subcontinental crust is the decisive tectonic factor. The rifting determined a slow development of the fold-and-thrust structure of the southern Kamchatka Peninsula. The general structural evolution progressed in a sedimentation basin and partly in periods of tectonic uplifts.

A stable orogenic structure has not been formed during the Cenozoic. In contrast, the inherited rifting of Miocene–Holocene period (in the continental crust practically) is accompanied by subalkaline and alkaline magmatism close to that of the Late Cretaceous–early Paleogene. Consequently, it can be assumed that Kamchatka did not reach yet the stage of typical island-arc development. The present-day island-arc systems are most likely products of an intricate magmatic and tectonic evolution.

ACKNOWLEDGMENTS

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The Tunaicha Formation and Its Age Analogues, Southern Sakhalin

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Abstract—The terrigenous Tunaicha Formation overlying dislocated Mesozoic basement was recently established at the base of Cenozoic deposits in southeastern Sakhalin (the Aniva and Dolinsk districts). The formation stratotype is in the Cape Tunaicha area (Bay of Mordvinov). The formation is attributed to the terminal middle–upper Eocene based on benthic and associated sparse planktonic foraminifers and on coeval floral assemblage. The Tunaicha period corresponded to formation time of new sedimentary basins in southeastern Sakhalin and Hokkaido.

Key words: South Sakhalin, Tunaicha Formation, stratotype, foraminifers, floral assemblage, spores, pollen, middle–upper Eocene.

INTRODUCTION

In the Aniva and Dolinsk districts of southeastern Sakhalin, predominantly coarse-grained volcanomictic deposits used to be attributed to the Arakai Formation of upper Oligocene, which represents the lowermost Cenozoic subdivision resting on complicatedly dislocated heterogeneous Mesozoic basement. The older Paleogene deposits have not been established here, although they are expectable by analogy with western Sakhalin.

Before recognition of the Arakai Formation, the deposits in question had been referred to various formations. Japanese geologists described them as the Morei and Aro (a part) formations. In nomenclature of L.B. Krivitskii, it was the Tomari Formation; I.G. Grinberg first attributed the rocks to the Khandasi and then the Gastello formations; I.N. Kuzina and L.S. Zhidkova, to the Gastello Formation, and finally, Tarasevich, to the Arakai Formation (*Geology...*, 1970).

New data on the basal Cenozoic strata of southeastern Sakhalin and their age were obtained by Zharov and Gal'versen (in 1993 and 1994) who compiled the regional geological map, scale 1 : 200000, and by Fot'yanova (in 1991), Serova, and Grokhotova (in 1993) who studied paleontological remains.

It was established that lower beds of the Arakai Formation are devoid of volcanogenic material. The beds are composed of sandstones, clay sandstones and siltstones intercalated with gritstone, conglomerate, and olistostrome lenses. Zharov and Gal'versen ascertained their wide distribution (Fig. 1).[1] In the Cape Tunaicha section, Fot'yanova discovered a floral assemblage dominated by the form genus *Macclintockia* (*M. pugetensis* Wolfe) that associated with diverse conifers, and mesophilic to subtropical flowering plants.

Benthic and scarce planktonic foraminifers found by Serova in the Cape Tunaicha (collection of Fot'yanova, 1991) and in the Aniva and Dolinsk districts (collections of Zharov and Gal'versen of 1993) suggest the terminal middle Eocene–late Eocene age of their host deposits.

Based on distinctive features of the basal Cenozoic deposits of southeastern Sakhalin (lithological diversity, wide distribution, and pre-Oligocene age), we discriminate them as the local Tunaicha Formation with the Cape Tunaicha stratotype section (Bay of Mordvinov).[2]

The Tunaicha Formation outcrops are known on eastern slopes of the Yuzhno-Kamyshovskii Ridge, in the western and southern structures surrounding the Susunaiskii Ridge, in the Korsakov Plateau and the Murav'evskaya Lowland (Fig. 1).

The Tunaicha Formation stratotype and some other sections are characterized in detail below along with the age substantiation.

† Deceased.

[1] Sheets of the State Geological Map of 1 : 200000 scale ("Geolkarta-2000"), South Sakhalin, were published not long ago (*State...*, 2001). The Tunaicha Formation is included into the map legend for the first time.

[2] The Tunaicha Formation is described herein for the first time, although the name was mentioned earlier (*Resolutions...*, 1998) without lithological–paleontological characterization and authorship; the unit age was defined imprecisely.

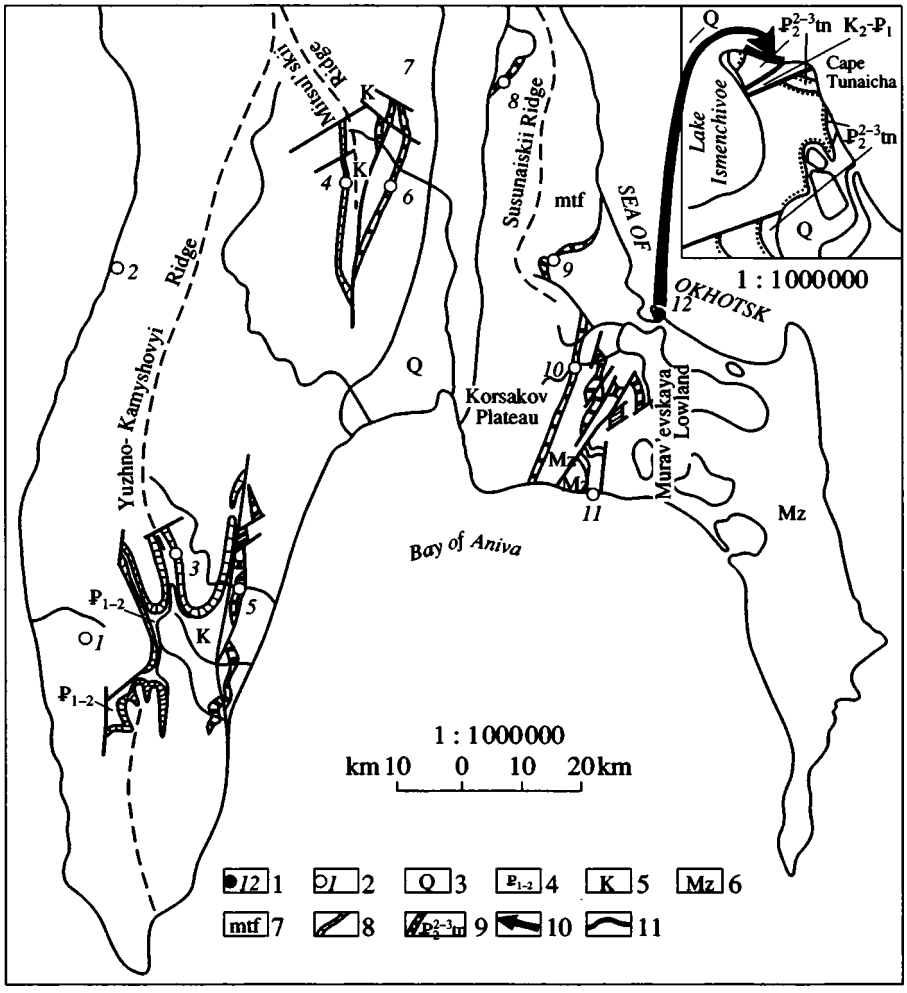

Fig. 1. Distribution area of the Tunaicha Formation in Sakhalin: (1) stratotype; (2) sections of (*1*) Kitosiya, (*2*) Pravda, (*3*) Vetvistaya, (*4*) Suchkovataya, (*5*) Uryum, (*6*) Yar, (*7*) Borehole 5-D, (*8*) Belaya, (*9*) Volynskaya, (*10*) Komissarovka, (*11*) Belokamennaya rivers and (*12*) Cape Tunaicha; (3) Quaternary deposits (Q); (4) Paleogene coal-bearing deposits (P_{1-2}); (5) undivided (K) and Upper Cretaceous (K_2) sediments; (6) Mesozoic siliceous–volcanogenic rocks (MZ); (7) metamorphic rocks (mtf); (8) outcrops of undivided Middle–Upper Paleogene Krasnopol'e–Takaradai formations; (9) outcrops of Tunaicha Formation (P_2 tn); (10) arrow pointing to the Tunaicha Formation stratotype area; (11) inset map of Tunaicha Formation stratotype area.

SECTIONS

The Aniva and Dolinsk Districts

Section 1. Cape Tunaicha, the formation stratotype (Figs. 1–4). The section is exposed along the Sea of Okhotsk coast, near the northern extremity of the Bay of Mordvinov, 950 km away to the west-northwest from the Cape Tunaicha extremity. The sea-cliff outcrops extend for a distance about 460 m. In the western slope of a large antiform, the formation beds dip westward 235°–240°, angle 20°–25°. Limbs of small asymmetrical auxiliary folds are 3 to 10 m wide, and hinge lines dip southwestward, angle 20°. A tectonic contact separates the Tunaicha Formation from undivided Lower Cretaceous deposits (Fig. 3). The succession beds above the contact is as follows:

Bed 1. Gray massive sandy siltstone grading into inequigranular clay sandstone. In the lower interval, N.M. Grokhotova identified *Ligodium japoniciformis, Ulmoideipites*, and *Tricolpites* characteristic of the Muz'ma palynological assemblage (Brutman, 1978, 1986). Apparent thickness is 30 m.

Bed 2. Light gray fine-grained sandstones with lens-shaped intercalations of gray fine-grained carbonate sandstones up to 0.5 m thick. Sample 1001/7 yielded foraminifers *Dentalina subsoluta* Cushm., *Pseudonodosaria charoensis* Asano, *Sigmoidella pacifica* Cushm. et Ozawa, *Cassidulina eobatula* Kaiho, *Eponides lobatus* Kaiho, *Melonis pompilioides* (Fichtell et Moll), and *Globigerina praeturritillina* Blow et Banner. Mollusks are represented by a single species *Venericardia* cf. *vagisana* Kogan. Fot'anova found *Cryptomeria* sp., *Liquidambar* sp., *Cinnamomum* sp., and *Macclintockia pugetensis* Wolfe (Locality 20/1 in

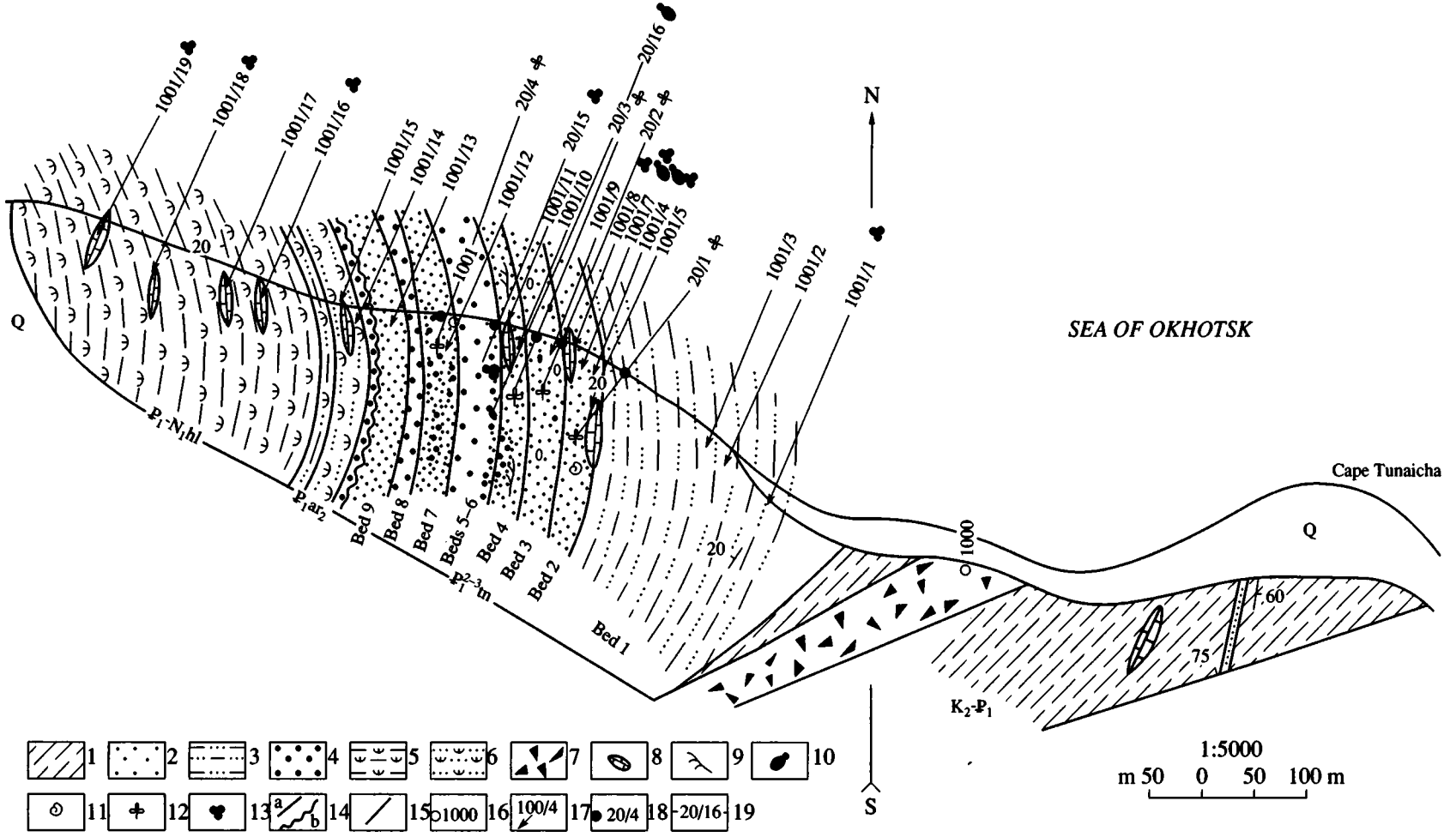

Fig. 2. Geologic scheme of the Tunaicha stratotype area, Cape Tunaicha (Mordvinov Bay): (1) mudstone; (2) sandstone; (3) clay sandstone; (4) gritstone; (5) tuffaceous siltstone; (6) tuffaceous sandstone; (7) olistostrome; (8) calcareous-clayey concretions; (9) plant detritus; (10–13) fossils: (10) foraminifers, (11) marine invertebrates. (12) macroflora, (13) spores and pollen; (14) boundaries of beds and formations: (a) conformable, (b) transgressive; (15) tectonic breaks; (16) observation points by Zharov and Gal'versen; (17) sites of sampling for microfauna and palynological remains; (18) sampling site of Fot'yanova; (19) microfauna site and sample no. Local stratigraphic units for Figs. 2–4: Middle Eocene, Krasnopol'e Formation ((P_2^2 kr); upper Eocene, Takaradai Formation (P_2^3 tk); terminal middle Eocene–upper Eocene, Tunaicha Formation (P_2^{2-3} tn); upper Eocene–lower Oligocene, lower Arakai Subformation (P_{2-3} ar_1); lower? Oligocene, upper Arakai Formation (P_3 ar_2); terminal Oligocene?–lower Miocene, Kholmsk Formation ($P_3 - N_1$ hl); upper Miocene–lower Pliocene, lower Maruyam Formation (N_{1-2} mr_1); other symbols as in Fig. 1.

Figs. 2–4). Grokhotova identified the Muz'ma palynological assemblage. Thickness is 5 m.

Bed 3. Light gray fine-grained, well sorted sandstone with small "floating" pebbles and chert gravel. Abundant leaf impressions of *Macclintockia pugetensis* Wolfe. Thickness is 5 m.

Bed 4. Gray fine-grained sandstone containing scattered siliceous gravel, big calcareous concretions, and abundant floral remains: *Pinus* sp., *Metasequoia occidentalis* (Newb.) Chaney, *Cryptomeria* sp., *Taiwania* sp., *Falcatifolium* sp. nov., *Magnolia* sp., *Cinnamomum* sp., and *Macclintockia pugetensis* Wolfe (Locality 20/3 in Figs. 2, 3). Thickness is 4 m.

Bed 5. Gray fine-grained sandstone with lenses and irregular nests of gravel. Thickness is 0.2 m.

Bed 6. Alternating layers (0.1–1.5 m) of gray fine-grained sandstone and gritstone with rare small pebble of metamorphic shales. Typical of gritstone layers are distinct, slightly wavy lower and gradual upper contacts. In the lower sandstones layers (Locality 20/16, collection of Fot'yanova, 1991), Serova identified foraminifers *Reophax tappuensis* Asano, *Haplophragmoides coalingensis* Cushm. et Hanna, *Haplophragmoides* sp. "B", *Haplophragmoides* aff. *yokoyamai* Kaiho, *Cyclammina incisa* Stache, *C. pacifica* Beck, *Matrinottiella* sp., "A", *Lenticulina antipoda* Stache, *Guttulina problema* d'Orb., *G. gibba globosa* (von Münster), *Sigmomorphina ezoensis* Kaiho, *Pseudoglandulina ovata* Cushm. et Applin, *Asterigerina crassaformis* Cushm. et Siegfus, *Osangularia wilcoxensis californica* (Mallory), *Melonis pompilioides* (Fichtell et Moll), and *Praeglobobulimina pupoides* (d'Orb.). At the base (Locality 20/15), Grokhotova detected a palynological assemblage that is similar in structure and composition to the Eocene Muz'ma assemblage (Brutman, 1978, 1986). Thickness is 6.5 m.

Bed 7. Gray fine-grained sandstone with thin (up to 0.1 m) gritstone lenses, abundant pelecypods preserved poorly and large leaf impressions of *Macclintockia pugetensis* Wolfe (Locality 20/4 in Figs. 2, 3). Thickness is 4 m.

Bed 8. Variegated massive gritstone. Thickness is 2.5 m.

Bed 9. Gray fine-grained massive sandstone. Thickness is 4.6 m. Apparent thickness of the Tunaicha Formation is 61.7 m.

Tuffaceous gritstones of the upper Arakai Formation overlie Bed 9 with a stratigraphic unconformity.

Section 2 (the Yar River section 6 in Figs. 1, 4) is exposed on the Mutsul'skii Ridge eastern slope, along the right tributary of the Vladimirovka River. The Tunaicha Formation rests on upper deeply eroded siltstone and mudstone beds of the Bykov Formation (Upper Cretaceous). The basal gritstone horizon (up to 2 m) includes pebbles of Cretaceous sandstone, siltstone, quartzite, and quartz–micaceous shale. The lower part of the formation is represented by gray fine- to

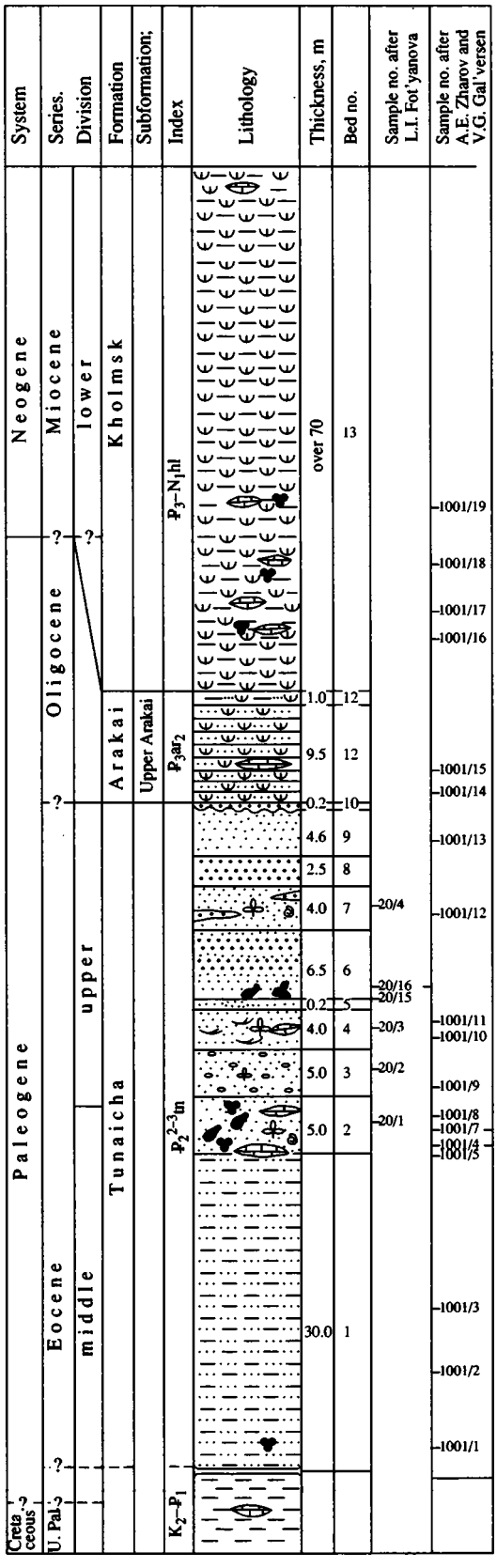

Fig. 3. Lithostratigraphy of the Tunaicha Formation stratotype, Cape Tunaicha (Bay of Mordvinov); symbols as in Fig. 2.

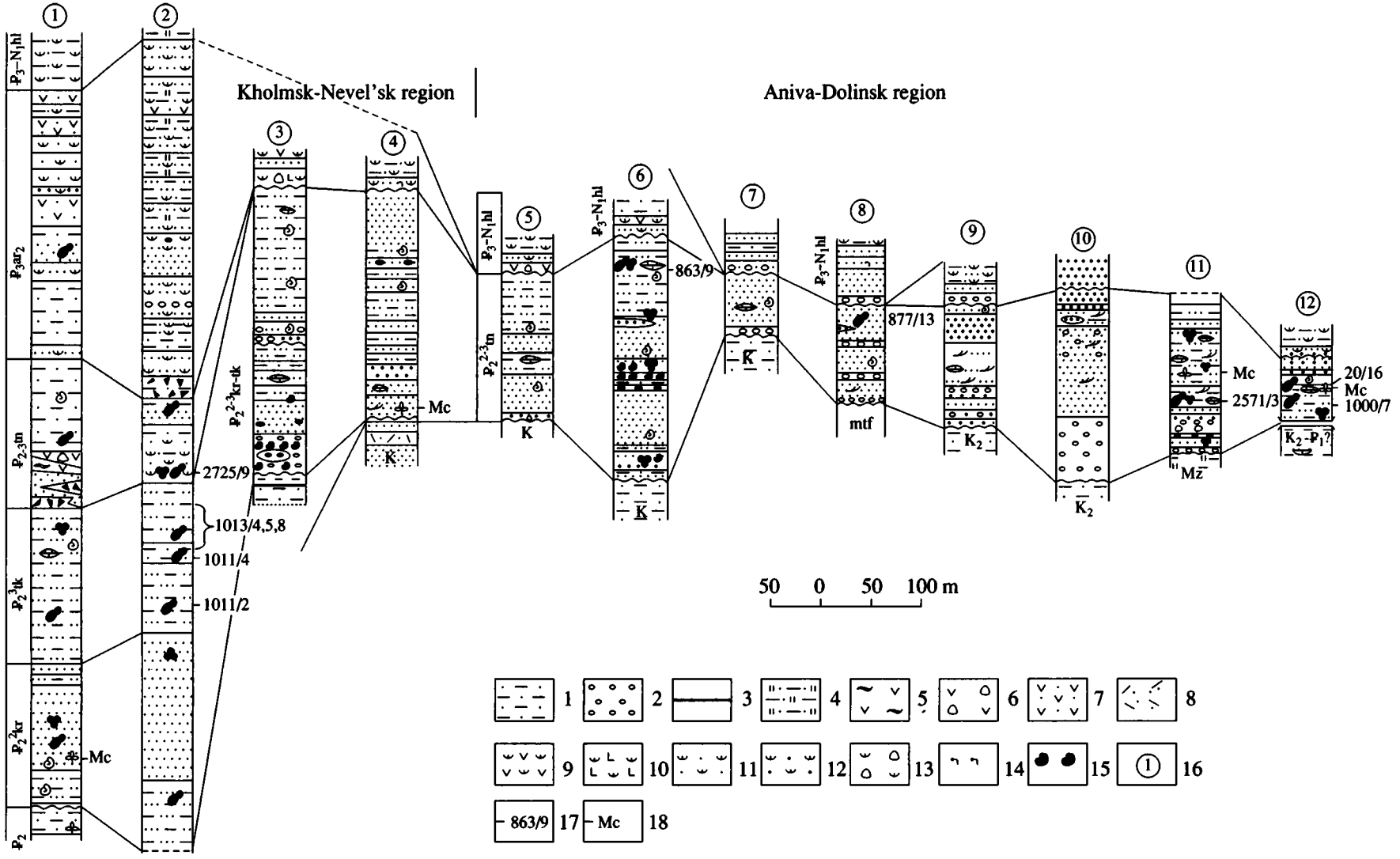

Fig. 4. Correlated sections of the Tunaicha Formation: (1) siltstone; (2) conglomerate; (3) coal seam; (4) siliceous siltstone; (5) lava breccia of intermediate composition; (6) agglomerate and (7) psammitic to silty tuffs; (8) aleuropelitic tuff of acid composition; (9–13) tuffite intermediate (9), basic (10), psammitic to silty (11), psephitic (12), and agglomeratic (13); (14) glauconite inclusions; (15) coquina interbed and lens; (16) section nos. as in Fig. 1; (17) samples with foraminiferal; (18) occurrence levels of *Macclintockia* (other symbols as in Fig. 2).

medium-grained sandstones with lenses and interbeds of gritstone (0.1–1.7 m), coquina (up to 0.7 m), rare siltstone and mudstone (up to 1.0 m) with minute lenses of sooty coals (up to 1 cm). Basal sandstones grade into upper fine-grained clay sandstone that yields rare foraminifers *Rhabdammina aspera* Volosh., *Haplophragmoides asagaiensis* (Asano), and *Perfectononion praeincertus* V. Kuzn. (Sample 863/9, Fig. 4). Mollusks *Yoldia matschigarica* Khom., *Venericardia (Megacardita?) matschigarica* Khom., *V. asagaiense* Mak., *Chlamys matchigarensis* Mak. and *Mya grewingki* Mak. (identifications of Zhidkova) are typical of coastal–marine and shallow-water ingression sediments (Zhidkova *et al.*, 1974). The spore–pollen assemblage from the formation lower part is similar to the Muz'ma assemblage of Bed 1 of the Tunaicha stratotype. In the Yar River section, the Tunaicha Formation is thicker (240 m) than elsewhere in southern Sakhalin.

Section 3 (the Belaya River section 8 in Figs. 1, 4) is exposed on the western slope of the Susunaiskii Ridge, in the Belaya River basin. The Tunaicha Formation rests on deeply eroded Upper Cretaceous deposits. A basal horizon of small-pebbled conglomerate is 3 m thick. Pebbles are predominantly (95%) derived from underlying sericite–chlorite para- and ortho-slate and quartzite. Conglomerate grades into gray homogenous fine- to medium-grained sandstone with interbeds and lenses of gritstone (0.3–0.8 m) and conglomerate (0.8–2.0 m). Sediments bear abundant thick-walled coastal–marine pelecypod shells and frequent remains of plant shoots. Rare foraminifers *Cyclammina tanai* Ishizaki and *Trochammina squamosa* Jones et Parker are found in sandstones (Sample 877/13, Fig. 4). Apparent thickness of the formation is 95 m.

Section 4 (the Belokamennaya River section 11 in Figs. 1, 4) is exposed in sea cliff of the Aniva Bay eastward of the town of Korsakov and Belokamennaya River mouth. The basal coarse conglobreccia consists of clasts and blocks up to 6 m across of diverse rocks: jaspers, highly siliceous sandstones, basic volcanics, tuffaceous sandstones, and tuffites of the underlying Utes sequence. The upper part consists of clay sandstone with interbeds of fine-grained sediments, numerous calcareous concretions with "gennoishi." Foraminifer taxa found in sandstones (Sample 2571/3 in Fig. 4) are *Bathysiphon eocenicus* Cushm. et Hanna, *Ammodiscus incertus* d'Orb. and abundant *Haplophragmoides* species *H. deflata* Sullivan, *H. asagaiensis* (Asano), *H. tanai* Kaiho, and *H. subevolutus* Kaiho. Secretory species *Nodogenerina sanctelcrucis* Kleinp., *Plectofrondicularia saitoi* Kaiho, *Eponides truempi* Nuttall, and *Asterigerina crassaformis* Cushm. et Siegf. occur as single specimens. The palynological assemblage is of the Kuz'ma type. In 1955 and 1956, geological team headed by Grinberg collected impressions of *Macclintockia*. Thickness of the formation is 145 m.

Section 5 along the Uryum River (Figs. 1, 4) was studied in the eastern Kril'on Peninsula, where the basal gritstone horizon (0.15–0.7 m) consists of angular clasts of Cretaceous sandstones and siltstones. Above it there are fine-grained and clay sandstones with unidentifiable plant fragments and mollusks *Clinocardium asagaiense* Mak., *Solen kovatschensis sachalinensis* Slod., *Mya grewingki* Mak., *Nemocardium ezoense* Takeda, and *Papyridea harrimani* Dall. (identified by V.P. Tuzov, GUP Dal'informgeocentr, Yuzhno-Sakhalinsk). Apparent thickness is 120–130 m. In the eastern part of the peninsula, the formation thickness ranges from 190 m in the north and down to zero in the south.

Some other sections shown in the correlation chart (Fig. 4) are briefly described below by the analysis of foraminifers (sections at the Kitosiya and Pravda rivers) and floral assemblages (sections at the Suchkobataya, Pereputka, and Obutonai rivers). Characteristics and distribution of foraminifers and floral assemblages in the Kitosiya section have been described earlier by Serova (*Geological...*, 1998; Fot'yanova *et al.*, 2001).

FORAMINIFERAL ASSEMBLAGES OF THE TUNAICHA FORMATION AND ITS AGE ANALOGUES

The Aniva and Dolinsk Districts

Foraminifers were found in the Tunaicha Formation stratotype area (Cape Tunaicha) and in the Yar, Belaya, Podorozhka, Gornaya, and Belokamennaya sections (Figs. 1, 4). Being of a low diversity, the foraminiferal assemblages consist of species characteristic mostly of the early Eocene in the North Pacific areas.

The most representative assemblage is that of the stratotype (locality 20/16). It consists mainly of agglutinated foraminifers *Bathysiphon eocenicus* Cushm. et Hanna, *Ammodiscus incertus* d'Orb., *Reophax tappuensis* Asano, *Cyclammina pacifica* Beck, *Martinottiella* sp. "A", and abundant *Haplophragmoides yokoyamai* Kaiho, *H. asagaiensis* (Asano), *H. deflata* Sullivan, and *Cyclammina incisa* Stache. These species dominate also in the assemblages from middle strata of the Poronai Formation of Hokkaido. More diverse secretory species are represented by single specimens of *Debtalina subsoluta* Cushm., *Pseudonodosaria charoensis* Asano, *Lenticulina antipoda* Stache, *Guttulina problema* d'Orb., *G. gibba globosa* (von Münster), *Sigmomorphina ezoensis* Kaiho, *Sigmoidella pacifica* Cushm. et Ozawa, *Pseudoglandulina ovata* Cushm. et Appl., *Eponides lobatus* Kaiho, *Asterigerina crassaformis* Cushm. et Siegf., *Osangularia wilcoxensis californica* Mall, *Cassidulina lobatula* Kaiho, and *Praeglobobulimina pupoides* d'Orb. *Melonis pompilioides* (Fichtel et Moll) occurs in abundance. Virtually all the listed benthic species found in the Aniva and Dolinsk areas are characteristic of the Poronai Formation in Hokkaido (Kaiho, 1984a, 1984b), the Kovacha Horizon of West Kamchatka (Tochilino and Korn formations), the Laparelam Formation of the Il'pinskii

Peninsula, Koryak Upland (Serova, 1967, 1969a, 2001) and the Karaginskii Island (Serova *et al.*, 1975).

The Tunaicha Formation yielded single planktonic foraminifers formerly unknown in Paleogene deposits of Sakhalin. *Subbotina praeturritilina* Blow et Banner that was found for the first time is characteristic of the basal upper Eocene (the Caribbean *Globigerapsis semiinvoluta* Zone). Along with the benthic assemblage, this species implies that enclosing deposits correspond in age to the initial late Eocene.

The upper part of the Tunaicha Formation in the Yar section (Sample 863/9) and its topmost part in the Belaya section (Sample 877/3) yielded only single agglutinating species *Rhabdammina aspera* Volosh., *Haplophragmoides asagaiensis* (Asano), *Cyclammina tanai* Ishizaki, *Trochammina squamosa* Jones et Parker, and *Perfectononion* sp. indet. All these species are of a wide stratigraphic range (the upper Paleogene of the North Pacific Province).

Foraminifers of Sample 2571/3 from the Belokamennaya section are more diverse and abundant. Like in the assemblage of the stratotype, they are dominated by agglutinated species, mostly by *Haplophragmoides deflata* Sullivan, *H. asagaiensis* (Asano), *H. yokoyamai* Kaiho, *H. tanai* Kaiho, and *H. subevolutus* Kaiho. Abundant agglutinating forms are associated with single secreting species *Nodogenerina sanctaecrucis* Kleinp., *Plectofrondicularia smithi* Kaiho, *P. packardi packardi* Cushm. et Schenck, *P. packardi multilineata* Cushm. et Sim., *Eponides truempi* Nuttall, *E. megapolitana* Massl., and *Asterigerina crassaformis* Cushm. et Siegf.

The Kholmsk and Nevel'sk Districts

Foraminiferal assemblages of the Tunaicha type were found in the Kitosiya and Pravda sections.

The Kitosiya section (Figs. 1, 4) is proposed to be the reference one for the southern structural–facies zone of the West Sakhalin depression (Fot'yanova *et al.*, 2001). Correlative late Eocene assemblages of foraminifers are known from the Krasnopol'e and Takaradai formations and from the lower Arakai Subformation. According to relative abundance and composition of species, these foraminifers are similar and even identical at some levels to their assemblages in the Poronai Formation of Hokkaido (Serova, 2001; Fot'yanova *et al.*, 2001). Lower strata of the Poronai Formation yield nannoplankton of CP14–lower CP15 zones (Okada and Kaiho, 1992) in association with benthic foraminifers. The lower and upper part of that formation belongs to the terminal middle and upper Eocene, respectively. Based on similarity of foraminiferal assemblages, above subdivisions of the Kitosiya section (Krasnopol'e, Takaradai formations and lower Arakai Subformation) can be attributed to the terminal middle–upper Eocene.

The Pravda section (Figs. 1, 4) of the Takaradai sandstone–mudstone sequence (about 350 m) is divided by Zharov into the lower sandstone–siltstone subformation with abundant tuffite material (about 320 m) and the upper subformation of tuffaceous sandstones (60 m). Above there are the Kholmsk (340 m) and Nevel'sk (apparent thickness about 80 m) formations of alternating siliceous tuffaceous sandstones, siltstones, and mudstones. Despite the frequent sampling, foraminifers have been detected only in the Takaradai Formation and lower Akarai Subformation.

The lower Takaradai strata (Sample 1011/2) yielded single specimens of planktonic *Globigerina incretacea* Chalilov unknown before in the Paleogene of Sakhalin. The species is characteristic of the upper middle Eocene (the upper part of the *Globigerina turcmenica* Zone) of the Crimean–Caucasian region. In the Far East, Serova described it from the upper Kilakirnun Formation of the middle Eocene (the upper part of the *Globigerina frontosa* Zone) in the Il'pinskii Peninsula, (Serova, 1969b; Krasheninnikov *et al.*, 1988).

The topmost part of the lower Takaradai Subformation (Sample 1012/4) includes scarce benthic foraminifers dominated by secreting forms. The most frequent is *Pseudoelphidiella khatyrkensis* Serova characteristic of the upper Eocene Ionai Formation of the Nizhne-Khatyrskaya depression (Serova *et al.*, 1977). It is associated with *Anomalinoides sasai* Kaiho, *Melonis lobatus* Kaiho, and *Caucasina eocaenica kamchatica* Serova. Two former species occur in the Poronai Formation of Hokkaido (Kaiho, 1984a, 1984b), and the latter subspecies is the index taxon of the synonymous zone in the Laparelam Formation of the Il'pinskii Peninsula (Serova, 1976). Agglutinated foraminifers are represented by single specimens of *"Discammina* sp. C" Kaiho and *Cyclammina* cf. *pacifica* Beck.

Species dominating in the upper Takaradai Subformation (Samples 1013/ 4, 1013/5, and 1013/8) are *Reophax tappuensis* Asano, *Haplophragmoides deflata* Sulliv., *H. tanai* Kaiho, *H. yokoyamai* Kaiho, *Cyclammina pacifica* Beck, *C. tanai* Ishizaki, *C. pusila* Brady, and *Trochammina* sp. Secretory forms are represented by *Pseudonodosaria* cf. *charlotensis* Asano, *Robulus pseudovortex* Cole, *Globobulimina pyrula* d'Orb., *Praeglobobulimina ovata* (d'Orb.), *"Rosalina* sp. A" Kaiho, *Gyroidina soldanii octocamerata* Cushm. et Hanna, *Cribroelphidium sumitomoi* (Asano et Murata), *C. sorachiense* (Asano), and *Melonis pompilioides* (Fichtell et Moll).

Abundant *Haplophragmoides apertiumbilicatus* Kaiho and *H. subevolutus* Kaiho characteristic of the Poronai Formation upper part were discovered at the top of the lower Arakai Subformation (Sample 2725/9). Thus, *Globigerina incretacea* Chalilov and associated benthic species of the *Caucasina eocaenica kamchatica* Zone, which are known from the Laparelam Formation of Il'Pinskii Peninsula and from the Poronai Formation of Hokkaido yielding nannoplankton of CP14–

CP16 zones, imply the terminal middle–late Eocene (late Bartonian–Priabonian) age of the lower Takaradai Subformation.

FLORAL AND PALYNOLOGICAL ASSEMBLAGES OF THE TUNAICHA FORMATION

Plant impressions occur sporadically in the Tunaicha Formation of the Aniva and Dolinsk districts and in its age analogues of the Kholmsk and Nevel'sk districts. They mostly represent the only form genus *Macclintockia* of flowering plants, largely the species *M. pugetensis* Wolfe (1968). A representative floral assemblage is recovered from beds 2–4 and 7 of the stratotype (Fig. 3). The most diverse plant remains are confined to Bed 4, whereas others contain *M. pugetensis* either alone (Beds 3 and 7) or in association with single species of conifers and flowering plants, such as *Cryptomeria* sp., *Liquidambar* sp., *Cinnanomum* sp., and others (Bed 2). In general, the fossil flora of the Tunaicha stratotype consists of diverse conifers populating various altitudinal belts of mountain forests. These are *Pinus* sp., *Metasequoia occidentalis* (Newb.) Chaney, *Cryptomeria* sp., *Sciadopitys* sp., *Taiwania* sp., and *Falcatifolium* sp. nov. Except for the latter occurring in abundance, other genera are represented by single specimens. Impressions of flowering plants *Magnolia* sp., *Liquidambar* sp., *Cinnamomum* sp., and others are also rare. *M. pugetensis* dominates in all localities.

During geological survey of 1955–1956, Grinberg collected impressions of *Macclintockia* in the deposits referred to the marine Gastello Formation at the Belokamennaya mouth.[3] It should be noted that flora buried in these deposits differs from that of the Gastello Formation stratotype (Gastello River basin). According to Shmidt (1989), the Gastello Formation yields there conifers *Metasequoia, Taxodium, Glyptostrobus* and flowering plants dominated by Betulaceae, while *Macclintockia* and other thermophilic genera of flowering and coniferous plants have not been found. The Gastello Formation flora is the late Oligocene in age (Shmidt, 1989), and deposits the Belokamennaya section cannot belong therefore to the Gastello Formation.

In the Kholmsk and Nevel'sk districts, *Macclintockia* impressions are discovered in four marine sections of the Suchkovataya, Pereputka, and Obutonai rivers and the Kholmsk area. In the Suchkovataya basin (pass area in the Kholmsk Mountains), Gal'versen collected in 1977 the *Macclintockia* impressions from the lower part of undifferentiated Krasnopol'e–Takaradai sequence (Fig. 4), which have been studied by Fot'yanova. In the Kholmsk area, *M. trinervis* Heer was discovered in the marine Nisi-Sakutan Formation (the Takaradai Formation according to *Resolutions...*, 1982) correlative with the Poronai Formation of Hokkaido (Krishtofovich, 1962). In the Pereputka and Obutonai sections (southwestern Kril'on Peninsula), plant impressions were collected in 1971 by Zhidkova and studied by Fot'yanova. According to Zhidkova, *M. pugetensis* Wolfe is buried in the Arakai Formation of the Pereputka section and in the Takaradai–Arakai boundary beds of the Obutinai section.

The Tunaicha Formation and its analogs containing *M. pugetensis* Wolfe correspond to the species acmezone (epibole) established by Fot'yanova at the middle–upper Eocene boundary in the Northwest Pacific biogeographic province and Alaska. The age established based on foraminifers by Serova is consistent with data on nannoplankton distribution (CP14b–CP15a zones, Okada and Kaiho, 1992). The *M. pugetensis* stratigraphic range can be wider however, because there is a site in Kamchatka (Karaginskii Island), where this species is confined to the uppermost Eocene level in the Peresheek Formation (Serova *et al.*, 1975) or, in the other nomenclature, to the top of the Cape Tons Formation (*Eocene...*, 1991). The *"Macclintockia"* floral horizon includes the following units: the Poronai Formation of Hokkaido, the Tunaicha Formation and correlative deposits of southern Sakhalin (the Krasnopol'e, Takaradai and lower Arakai formations of the Kitosiya section), the Tochilino and Korn formations of central West Kamchatka. The correlative stratigraphic subdivisions of Alaska are the Katalla Formation basal beds and upper parts of the Kushtaka and Kulthieth formations. In the State Washington, the horizon corresponds to the Renton Formation and unnamed volcanogenic sequences overlying the Puget Formation. Wolfe (1968, 1977) referred the listed units of North America to the provincial Kummerian Stage.

The Muz'ma palynological assemblage was identified by O.V. Vysochina in Bed 1 and by N.M. Grokhotova in Bed 6 of the Tunaicha stratotype. It is comparable in composition with the Eocene Muz'ma assemblage of the Machigar Horizon of Sakhalin (Brutman, 1986). The assemblage is composed of abundant ferns, conifers dominated by Pinaceae, thermophilic *Podocarpus* and *Dacrydium*, diverse angiosperms, Paleophytic form genera, and tree-ferns of the Dicksoniaceae and Schizaceae families.

EOCENE SEDIMENTATION, PALEOTECTONIC AND PALEOGEOGRAPHIC SETTINGS OF SOUTHERN SAKHALIN

It was the Eocene time when not only southern Sakhalin but also the entire Hokkaido–Sakhalin fold zone experienced significant structural changes, which transformed the Asia–Pacific Ocean transition zone. A new system of Cenozoic sedimentary sea basins appeared in the Hokkaido–Sakhalin region.

The comparative lithological–facies analysis of the Tunaicha Formation and coeval well-studied Kras-

[3] Identification of Fot'yanova is based on drawing by L.S. Zhidkova from the Russian Research Geological Prospecting Institute, St. Petersburg.

nopol'e, Takaradai, and Arakai (lower part) formations shows that the middle–late Eocene sedimentation was sharply different in the eastern and western areas of southern Sakhalin. As it is seen in latitudinal profiles, the western lithological facies give place to the eastern ones in eastern spurs of the West Sakhalin Mountains, where changes in composition and thickness of deposits and in age and composition of the underlying substratum are recorded. In terms of structural geology and tectonics, the boundary between the western and eastern structural–lithological zones corresponds to the Central Sakhalin fault zone traceable in axial areas of the Early Cenozoic Kamyshevskii and Kril'on uplifts with Cretaceous deposits exposed in their arched parts (Fig. 5). In general, this boundary separates the western front of folding from area of the Early Paleogene ascending movements in southwestern Sakhalin.

The formation of the Cenozoic sea basins was preceded by the Paleocene closure and inversion of the Cretaceous Iezo–West Sakhalin fore-arc basin and by development of marginal epicontinental coal-bearing (Snezhinka) sedimentary basin in the early Eocene. This basin extends for 600 km along western Sakhalin, from the Kril'on Peninsula in the south to the Aleksandrovsk-Sakhalinskii area in the north. It is located in front of the accretion–collision structures of eastern Sakhalin, which were formed at the same time (Paleocene–early middle Eocene, particularly in southeastern Sakhalin) as a result of collision and deep-seated underthrusting of heterogeneous blocks, which experienced metamorphism of a moderate pressure (Kiminami *et al.*, 1992; Bazhenov *et al.*, 2002; *State...*, 2001). There could be a dissected mountainous area favorable for intensive erosion of pre-Cenozoic deposits, from where clastic material was transported into the Snezhinka depression. This is evident from absence of upper Paleocene–basal middle Eocene sediments in southeastern Sakhalin, from uniform composition of pebbles (greenstones, cherts) in conglomerates of the Snezhinka Formation, and from close ages of granitoids in the northern Tonino-Aniva Peninsula and youngest metamorphic rocks of the Susunaiskii Ridge (*State...*, 2001).

The middle–upper Eocene deposits of western (Krasnopol'e-Takaradai) and eastern (Tunaicha) sections of southern Sakhalin recorded the onset of the Paleogene transgression. In the typical western sections (Kitosiya, Pravda, Savidskaya, Storozhka and other rivers), sediments overlie conformably or unconformably the subcontinental coal-bearing Snezhinka Formation of the Lower Paleogene. In the eastern sections, they rest unconformably, with a considerable local hiatus on heterogeneous basement composed of Triassic–Jurassic jaspers, Cretaceous olistostromes and turbidites, Cretaceous–Paleocene metamorphic slates, and of Upper Cretaceous and Paleocene–lower Eocene tectonic melanges. In all western sections, the basal coastal–marine and shallow-water sandy deposits of Krasnopol'e Formation grade into the marine clayey (Takaradai Formation) and tuffaceous sandy–clayey sediments of the lower Arakai Subformation (Fig. 4, columns 1, 2). In contrast, sections of the Tunaicha Formation (Fig. 4, columns 6–11) show transition from dominating coarse clastic deposits and olistostrome beds coastal–marine glauconitic sandstones with thin clay members. An exception is the easternmost sections of the northern Tonino-Aniva Peninsula, where basal conglomerates are abruptly substituted by sandy-clayey, frequently flyschoid and tuffaceous marine sediments.

The distribution patterns and structure of the Tunaicha Formation section suggest that sedimentary basin of that time represented a regular system of troughs and semi-isolated depressions of northeastern orientation, which appeared after destruction of mountainous area in southeastern Sakhalin (Fig. 5). The troughs were surely connected with the Krasnopol'e–Takaradai basin in the west and faced a deeper sea basin in the east; slope facies of that basin are discovered in the northern Tonino-Aniva Peninsula. The Krasnopol'e–Takaradai basin was separated from the Tunaicha basin by the Kamyshevskii and Kril'on arcs of the volcanism-free (structural) type.

In general, the Krasnopol'e–Takaradai basin inherited orientation and structural peculiarities of the marginal Snezhinka depression, which had originated at the collision time being bordered in the west by an active volcanic zone (the Moneron Rise). The volcanic zone extended from central part of the Sea of Japan (the Yamato Rise) and was, according to petrochemical data, comparable with structures of the mature pericontinental island arc.

Thus, the western Krasnopol'e–Takaradai basin was definitely of the inter-arc type. Position of the Tunaicha basin located easterly is less definite. The structure of Tunaicha sections and the eastward transition to deeper sea facies imply that sedimentation of the Tunaicha time developed in an outer inter-arc basin bounded in the east and southeast by a fore-arc terrace or a complicatedly terraced upper slope of a deep trench. The trench was likely located to the east, in the recent Yuzhno-Okhotsk depression. The assumption is consistent with occurrence of upper Eocene hemipelagic sediments and turbidites in the northeastern part of central Hokkaido (the Monbetsu area), which were evidently deposited in lower settings of the trench slope (Kiminami *et al.*, 1992). Morphologically, the Tunaicha basin represented a shallow sea with an archipelago that was connected with deeper basins in the east and west.

Data considered above suggest that southern Sakhalin of the Eocene time was a segment of typical active continental margin of eastern Asia, where a system of volcanic and non-volcanic arc-like rises was separated by troughs. In the Late Cretaceous–Early Paleogene, the continent–ocean transitional zone extended along

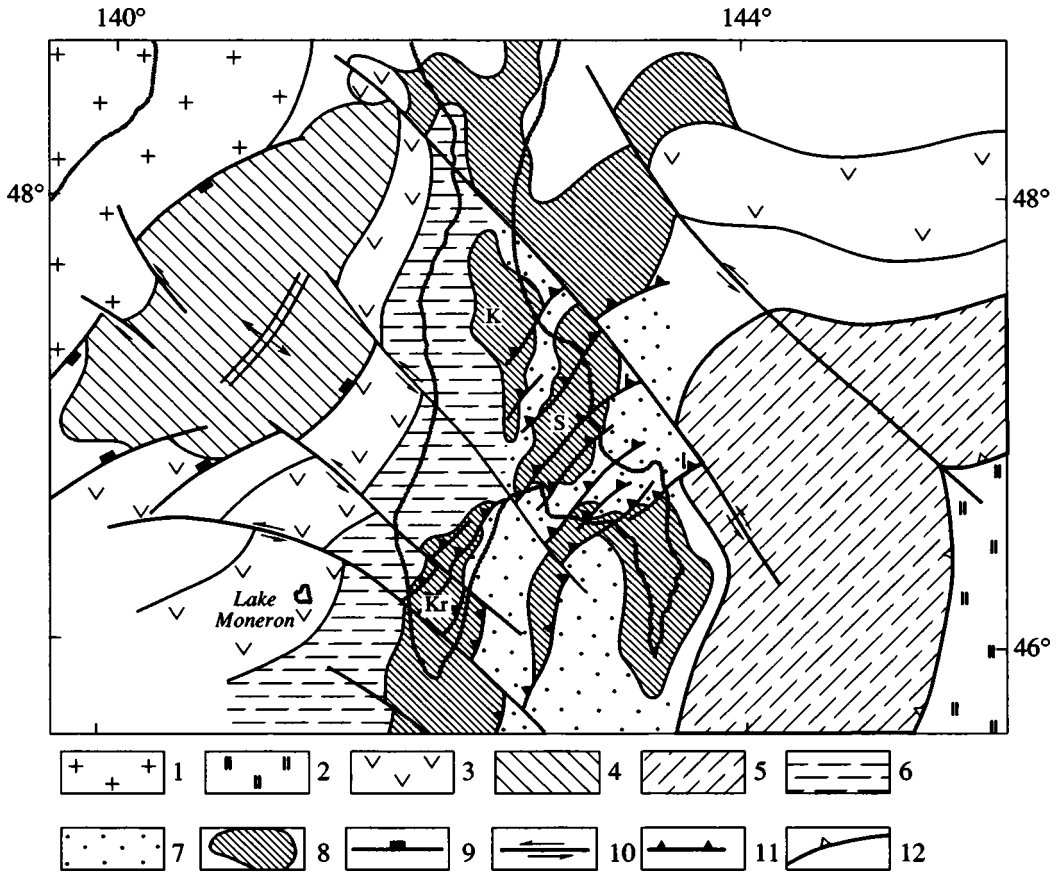

Fig. 5. Paleoectonic map (terminal middle–late Eocene) of the southern Sakhalin segment of the continent–ocean transition zone near eastern Asia (modern coordinates; with the Miocene opening of the Tatarskii Strait and the late Miocene-Quaternary movements of Sakhalin along dextral slip faults are taken into consideration): (1) Asiatic continental margin; (2) oceanic plate; (3) volcanic island arc; (4) back-arc basin; (5) fore-arc basin, deep trench slope; (6) inner Krasnopol'e–Takaradai and (7) outer Tunaicha inter-arc basins; (8) structural rises of (K) Kamyshevoe, (Kr) Kril'on, and (S) Susunaiskoe non-volcanic arcs; (9) fault, (10) shear, (11) upthrow and thrust faults; (12) deep trench axis.

the entire length of Sakhalin, whereas in the Eocene time, after the structural rearrangements, it changed orientation, at the latitude of southern Sakhalin, from the submeridional northeastern to sublatitudinal and extended north-northeastward toward Kamchatka. The Tunaicha sedimentary basin was therefore at the bending point of frontal fore-arc elements of the past active margin.

There are no analogs of the Tunaicha Formation northward of its distribution area described in this work (Fig. 1). Most likely, they must not be there, because the area was occupied by the Gastello sedimentary basin of the inter-arc type.

The Krasnopol'e, Takaradai and Arakai formations of the West Sakhalin structural–facies zone are of different stratigraphic ranges in southern and central parts of Sakhalin. In the Kril'on subzone (Pravda and Kitosiya river basins) they accumulated during the late middle–late Eocene, while northward, in the Uglegorsk district (Aral'skaya river basin), the facies analogues of Takaradai and Arakai formations were deposited in the cold Oligocene time (Serova, 1985). This is evident from comparative analysis of foraminiferal assemblages and from absence of thermophilic *Macclintockia* remains in northern sections of Sakhalin, the appearance of which in northern latitudes of Pacific province is recorded in the terminal middle–initial late Eocene.

Diachronism of boundaries separating sequences (formations) of different lithofacies affinity is understandable in view of data implying progradation of the Paleogene transgression from the south to the north within the Sea of Japan region.

CONCLUSION

The Tunaicha Formation is an individual local stratigraphic subdivision. Its terrigenous sediments lacking admixture of volcaniclastic material are widespread in southeastern Sakhalin.

Based on planktonic foraminifers, Serova concluded that the Tunaicha Formation range corresponds

to the terminal middle–late Eocene, i.e., to the late Bartonian–Priabonian interval of the European scale. According to floral evidence (*Macclintockia* assemblage), the formation accumulated during the middle–late Eocene that was a time of *M. pugetensis* Wolfe epibole in the Pacific biogeographic province. Characteristic of the formation is also the Eocene palynological assemblage of the Muz'ma type.

The Tunaicha Formation is correlative with the Krasnopol'e and Takaradai formations and lower Arakai Subformation of the Kril'on Peninsula, southwestern Sakhalin, and with the Poronai Formation of Hokkaido.

The Tunaicha time marked destruction of the Mesozoic thrust-and-fold basement. Simultaneously, it was the formation time of new Cenozoic sedimentary basins in southeastern Sakhalin and Hokkaido.

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The Age of the Baraba Formation Inferred from the U/Pb (SHRIMP) Dating (Sredinnyi Range, Kamchatka): Geological Consequences

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Abstract—The Baraba Formation exposed in the Sredinnyi Range (Kamchatka), where it is composed of terrestrial conglomerates, was considered for a long time as the late Campanian–lower Maastrichtian in age based on the fossil flora evidence. Zircons from the tuff layer in the lower part of the formation are dated at 50.5 ± 1.2 Ma using the U/Pb (SHRIMP) method, i.e., the accumulation of their host deposits commenced in the terminal early Eocene. The Baraba Formation overlies unconformably schists and phyllites of the Malka Group, which were metamorphosed during the pre-middle Eocene time, and the Irunei marginal-sea sediments thrust over metamorphic complexes and thus suggesting the pre-middle Eocene thrusting stage.

Key words: Eocene, zircon, U/Pb SHRIMP, Baraba Formation, Sredinnyi Range, Kamchatka.

INTRODUCTION

The present-day structure of the Kamchatka Peninsula was formed in the course of accretionary–collision processes that occurred in the terminal Mesozoic–Cenozoic in the northwestern margin of Eurasia (Bogdanov and Chekhovich, 2002; Konstantinovskaya, 2002). One of the most important events in the Cenozoic regional history was collision of the Cretaceous island arc with Eurasia (Konstantinovskaya, 2000; Soloviev *et al.*, 2002). The northern segment of the arc was thrust over the continental margin (the Lesnaya suture) in the middle Eocene (Soloviev *et al.*, 2002). Its southern (Valaginskii) segment is trust along over metamorphic rocks (the Andrianovka suture) of the Kamchatka Sredinnyi Range that are considered as the exhumed part of the Okhotsk continental block (Khanchuk, 1985; Konstantinovskaya, 2000) or the uplift of the West Kamchatka microplate basement (Bogdanov and Chekhovich, 2002). From the east, metamorphites of the Sredinnyi Range are overridden by the Cretaceous–Paleogene rock complexes that were formed within the Valaginskii island arc and marginal sea separating the arc from the Eurasian margin (Zinkevich *et al.*, 1994; Konstantinovskaya, 2002).

The Baraba Formation exposed in the Sredinnyi Range of Kamchatka and composed of terrestrial conglomerates was considered for a long time as an oldest neoautochthon because it overlies unconformably both the metamorphic rocks and Cretaceous marginal-sea sediments (Shapiro *et al.*, 1986). Based on the floral remains (determination by A.I. Chelebaeva), the Baraba Formation is dated back to the late Campanian–early Maastrichtian (Shapiro *et al.*, 1986). Data on the formation age, its structural position and composition have been used to construct geodynamic models of the southern Kamchatka evolution (Shapiro *et al.*, 1986; Zinkevich *et al.*, 1994; Rikhter, 1995; Konstantinovskaya, 2000). In this work, we consider new data on the U/Pb (SHRIMP) age of zircons from basal layers of the Baraba Formation and some geological consequences inferable from dating results.

THE SREDINNYI RANGE STRUCTURE

Metamorphic rocks exposed in the southern part of the Sredinnyi Range occur within a belt approximately 200 km long and 30–40 km wide, extending in the meridional direction (Fig. 1). They are subdivided into three structural complexes. The lower one corresponds to the *Kolpakova Group* (highly metamorphosed migmatized gneisses, gneisses, crystalline schists) intruded by two-mica granites of the *Krutogorova Complex.* This structural stage was considered as a nucleus (or basement) of the Kamchatka median mass (Khanchuk, 1985; Rikhter, 1995). The age interpretation of the Kolpakova metamorphites and Krutogorova granites is ambiguous (*Explanatory Notes...*, 2000). The Kolpakova and Krutogorova complexes are unconformably overlain, locally with basal conglomerates, by metamorphosed terrigenous rocks of the *Shikhta Formation* (Khanchuk, 1985; Rikhter, 1995). The latter is composed of garnet–biotite–staurolite schists, staurolite–

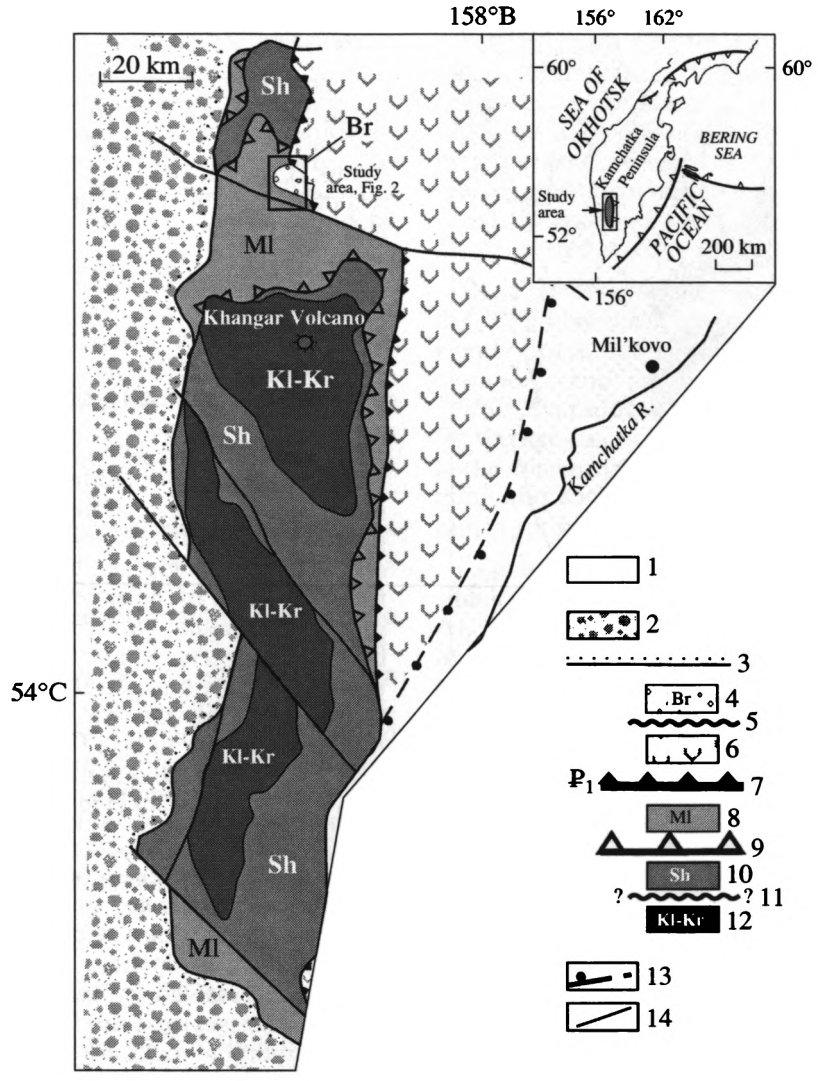

Fig. 1. Simplified tectonic map of the Sredinnyi Range, Kamchatka (Zinkevich *et al.*, 1998, with modifications) and schematic relationships between principal structural units: (1) Miocene–Quaternary sediments of the Central Kamchatka graben; (2) Eocene–Quaternary sediments of the West Kamchatka basin; (3) transgressive overlapping; (4) neoautochthon (Baraba Formation (Br), 50 Ma; (5) unconformity of the base of the Baraba Formation (Shapiro *et al.*, 1986); (6) Upper Cretaceous rocks of the island-arc terrane; (7) Andrianovka nappe; (8) Malka complex (Ml); (9) thrust fault (Rikhter, 1995); (10) Shikhta (Kamchatka) complex (Sh); (11) unconformity of the base of the Shikhta Complex (Khanchuk, 1985; Rikhter, 1995); (12) Kolpakova and Krutogorova complexes (Kl+Kr); (13) normal faults; (14) subsidiary faults.

sillimanite and biotite–muscovite plagiogneisses, and migmatites (Khanchuk, 1985; Rikhter, 1995). The *Malka Complex* that includes the Andrianovka, Kheivan, and Khimka formations is thrust over the Shikhta Formation (Rikhter, 1995). The Andrianovka Formation of amphibole, epidote–amphibole, and clinopyroxene–amphibole crystalline schists and amphibolites was regarded as correlative with the Alistor Formation (Bondarenko, 1997) and conformably overlain by the metamorphosed terrigenous Kheivan Formation, which is replaced higher in the section by the metamorphosed volcanics of the Khimka Formation. Rocks of the Malka Group are metamorphosed in the greenschist to amphibolite facies (Khanchuk, 1985).

In the western slope of the Sredinnyi Range, the unmetamorphosed Upper Cretaceous Khozgon Formation is composed of flyschoid quartz–feldspar sediments (Shapiro *et al.*, 1986) and has tectonic contacts with metamorphic complexes.

The Upper Cretaceous Irunei and Kirgakin formations of the Irunei nappe are thrust (the Andrianovka suture) over both the metamorphic complexes and the Khozgon Formation. The Irunei Formation (Santonian?–Maastrichtian) consists of terrigenous–siliceous–volcanogenic sediments accumulated in the marginal-sea basin and on the island-arc slope. The Kirgakin Formation (upper Campanian?–Maastrichtian) is composed of rudaceous tuffaceous rocks and

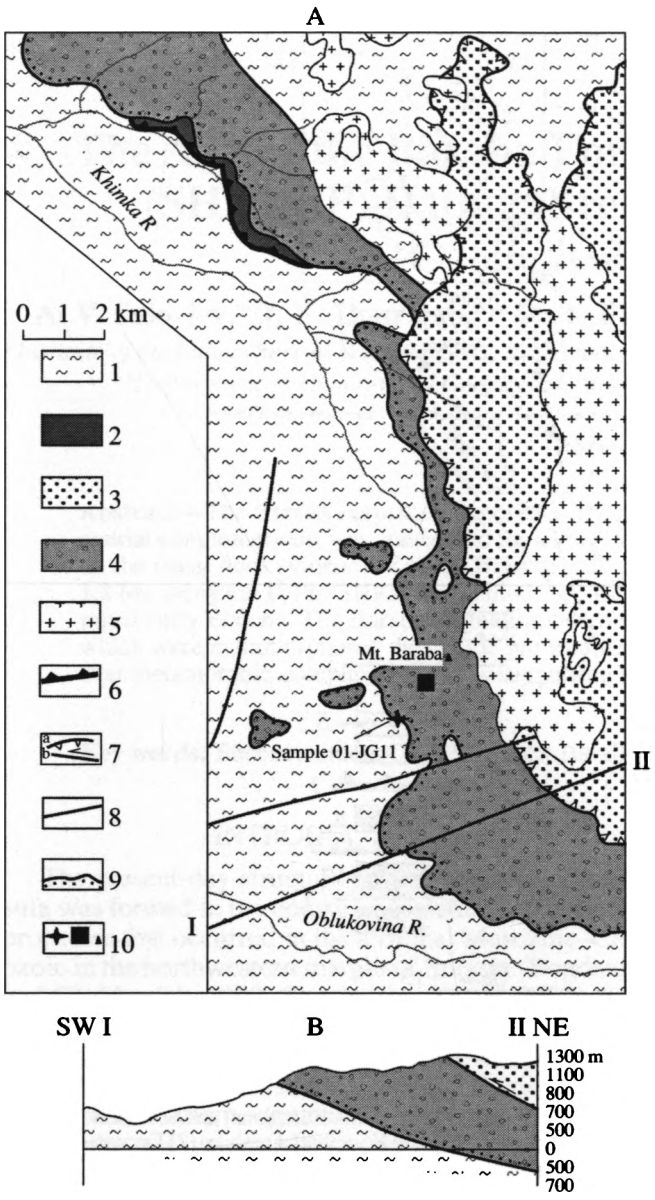

Fig. 2. Schematic geological structure of the Mt. Baraba area (Sredinnyi Range, Kamchatka) (A) and schematic cross section along line I–II (B) (Slyadnev *et al.*, 1997, modified based on materials from Shapiro *et al.* (1986) and Zinkevich *et al.* (1994)): (1) undivided metamorphic rocks of the Malka Group and Shikhta Formation of problematic age; (2) Upper Cretaceous tuffaceous–siliceous Irunei Formation; (3) Upper Cretaceous terrigenous Khozgon Formation; (4) Eocene tuffaceous–siliceous Baraba Formation; (5) Miocene granites; (6) thrust fault separating Irunei Formation from overridden metamorphic rocks; (7) thrust fault separating Khozgon Formation from overridden Baraba Formation in the map (a) and cross section (b); (8) subvertical faults; (9) unconformable boundary; (10) asterisk designates sampling sites for U/Pb dating, square shows flora locality, and triangle corresponds to Mt Baraba summit.

basic to intermediate volcanics formed in the island-arc system (Zinkevich *et al.*, 1994).

The neoautochthon above metamorphic rocks and the Irunei allochthon are represented by the Baraba Formation, the detailed description of which is given below (Fig. 2). The unconformable stratigraphic contact of the Baraba sediments with metamorphic and Irunei rocks was described by I.A. Sidorchuk and E.M. Ereshko in the course of geological mapping and thematic works (Shapiro *et al.*, 1986).

GEOLOGIC POSITION AND CHARACTERISTICS OF THE BARABA FORMATION

In its stratotype area (Mt. Baraba area, Fig. 2), the Baraba Formation is subdivided in two sequences. The lower sequence unconformably overlying schists and phyllites of the Malka Group is composed of tuffaceous and volcanic conglomerates, tuffstones, and andesitic tuffs. Pebbles of hornblende-dacite porphyrite, porphyry-like granite, basalts, acidic tuffs, tuffaceous siliciliths, jasper, cherts, and sandstones are cemented by sandy pyroclastic material similar in general composition to the pebbly material. The rock fragments are well rounded, ranging in size from sand grains to boulders. The lower sequence is approximately 150 m thick (Shapiro *et al.*, 1986; Zinkevich *et al.*, 1994).

The upper sequence conformably overlying the lower one is composed of polymictic conglomerates, breccia–conglomerate, gravelstones, sandstones, and subordinate siltstones. Cement in the rocks is sandy, sometimes aleuropelitic. Coarse clastic material is largely represented by metamorphic rocks (phyllites, schists, gneisses) and subordinate volcanics. Fragments of metamorphic rocks are subangular, unsorted, associated with better rounded granitoid, quartz, and feldspar pebbles derived from a more remote provenance (Kolodyazhnyi *et al.*, 1996). The upper sequence is approximately 800 m thick.

The stratigraphic boundary between the lower and upper sequences is sharp, probably implying a rapid change of provenances. Although there are no transitional varieties with equal proportions of volcanogenic and metamorphic material, the upper part of the lower sequence contains abundant fragments of metamorphic rocks, while basal layers of the upper sequence enclose fragments of volcanic rocks (Shapiro *et al.*, 1986).

In the right bank of the Kapitanskaya River 1.5 km of its mouth, there is exposed a tuffaceous–siliceous unmetamorphosed section 10 m thick, which we refer to the Irunei Formation. The latter overlies chlorite schists and phyllites of the Kheivana Formation.

The Baraba Formation is composed of continental facies. The fossil floral assemblage from the southwestern slope of Mt. Baraba is determined to be the late Campanian–early Maastrichtian in age (Shapiro *et al.*, 1986).

Pebbles from the Baraba Formation were studied in thin sections. Pebbles of volcanic rocks correspond in composition to rhyolites, dacites, andesites, basaltic andesites, and basalts. Pebbles of basic and intermediate volcanics could be derived from the eroded Mesozoic Irunei, Kirganik, and Kvakhona formations (Zinkevich *et al.*, 1994; Konstantinovskaya, 2002) and those of acidic and intermediate volcanics could be related in origin with the Paleocene Cherepanova Formation (Slyadnev *et al.*, 1997) and Eocene rocks of the Mt. Chernaya area (Gladenkov *et al.*, 1997). Pebbles of metamorphic rocks are identical in composition to rocks from the Shikhta Formation and Malka Group, and some of them are composed of gneisses similar to those form the Kolpakova Group. Granite pebbles could originate from the Krutogorova Complex (Khanchuk, 1985; Rikhter, 1995).

A comparison of sandstone pebbles from the Baraba conglomerate showed that they are compositionally identical to rocks of the Khozgon Formation. Radiolarites (cherts, jaspers) from pebbly fraction appeared to be similar to their counterparts from the Irunei Formation. Thus, the rocks of both the past Eurasian margin (metamorphites, granites of the Krutogorova Complex, sandstones of the Khozgon Formation) and the collided island arc (volcanics and radiolarites of the Irunei Formation) are present among pebbles of the Baraba conglomerates. The mixed pebbles derived from both the autochthon and allochthon suggest that the Baraba Formation formed after tectonic juxtaposition of two structural units.

THE BARABA FORMATION AGE

The closure temperature of U/Pb system in zircons is estimated to be over 900°C (Lee *et al.*, 1997). It is believed that the zircon age measured by U/Pb method determines the crystallization time, as the U/Pb system of this mineral is very resistant to thermal impacts.

Following the standard procedure, zircon crystals were extracted at the Institute of the Lithosphere of Marginal Seas from dacite tuff sampled from the basal layer of the Baraba Formation (Sample 01JG-11). Approximately 50 zircon grains were picked manually from the sample. Zircons from the sample and standard AS57 (Paces and Miller, 1993) were set into epoxy resin and then polished. The absence of fissures and inclusions in zircon grains was checked under magnification of 20 in the reflected and transmitted light. To study zoning and internal structure of polished zircons, we used scanning electron microscope JEOL JSM 5600 equipped with the cathode–luminescence detector (Fig. 3, A). The cathode–luminescence images revealed that zircon grains are lacking xenogenous nuclei. All the grains are prismatic and euhedral. The CL-zoning is characteristic of zircons from igneous rocks.

The isotopic ratios were measured using the Sensitive High-Resolution Ion MicroProbe–Reserve Geometry (SHRIMP-RG) equipment at the Stanford-USGS Microanalytical Center following the standard technique (Muir *et al.*, 1996). The beam of negatively charged oxygen atoms approximately 30 μm across was used for ionization of the analyzed crystals. Each measurement consisted of five cycles. After four to five runs in crystals of unknown age, we did measurements in the age standard AS57. The U and Th concentrations are calibrated relative to SL13 (Williams, 1998).

Age values presented in the table are obtained after the ^{207}Pb-correction based on assumption that insignificantly discordant zircons contain a simple mixture of common and radiogenic lead. The measured $^{207}Pb/^{206}Pb$ ratios are used to correct the common lead. By the age calculation, we extrapolated measured data to the concordia line of the model common lead (Cumming and Richards, 1975) to approximate ages determined for individual grains.

Zircons from Sample 01JG-11 are represented by small (±50–150 μm) zoned euhedral crystals (Fig. 3, A). Based on the weighted average ^{207}Pb value corrected for $^{206}Pb/^{238}U$ ages of 15 grains, the calculated age is equal to 50.5 ± 1.2 (±2σ) Ma (Fig. 3, B). The MSWD value of 0.90 shows that the data scatter does not exceed the analytical inaccuracy. A discordant age of one grain was omitted in calculations of the mean-weighted age.

Thus, the crystallization age of zircons from tuffs of the Baraba Formation is 50.5 ± 1.2 Ma. Inasmuch as tuffs formed instantly in terms of the geological time, the age of zircons is close to that characterizing the accumulation time of the Baraba Formation lower layers, and overlying conglomerates should be consequently younger than 50 Ma.

DISCUSSION

A.V. Shcherbakov was first to define the Baraba Formation prior to the World War II, although this subdivision received the official status later on (*Geology of the USSR*, 1964). Only plant remains are known to characterize sediments of the formation. During the geological survey, Yu.V. Makarov, M.I. Goryaev, and Yu.S. Voronkov sampled floral remains. Yu.V. Shtempel' and L.Yu. Budantsev who examined sampled flora referred the Baraba Formation to the Eocene (probably, Paleocene?–Eocene). B.M. Shtempel' correlated the Baraba flora with the floral assemblage from the lower layers of the Napan Formation in the Omgon Cape area (*Stratigraphy of the USSR*, 1975). Later on, M.N. Shapiro with colleagues performed additional sampling at the southwestern slopes of Mt. Baraba. A.I. Chelebaeva, who examined these remains, arrived at the conclusion that the Baraba flora is the late Campanian–early Maastrichtian in age (Shapiro *et al.*, 1986) based on the *Nilssonia* representatives present in the floral assemblage. Nevertheless, according to the private communication of M.A. Akhmet'ev (Geological Insti-

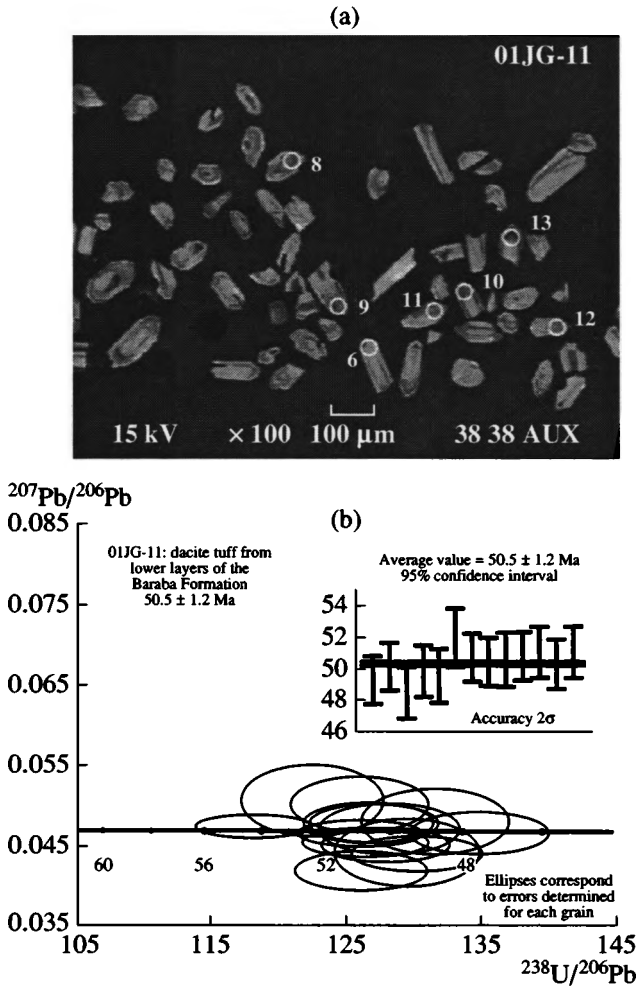

Fig. 3. (A) Cathode luminescence images of zircon crystals from Sample 01JG–11 with numbers of dated grains and white circles indicating points of SHRIMP analyses with analysis numbers shown in the table. (B) The Tera-Wasserburg plot for Sample 01JG-11, where horizontal solid line is concordia; the weighted-average age value of 50.5 ± 1.2 Ma (±2σ) is calculated based on fifteen grains and one with significant admixture of common lead (see parameters of discordance in the table) is excluded from calculations.

tute, Russian Academy of Sciences), some cycadophytes or ferns, such as *Dion* or *Cycas*, which are known from the Paleocene and Eocene sediments of Alaska, Japan, and northeastern China, can be mistaken for *Nillsonia* in case when remains are poorly preserved. In addition, the floral assemblage studied by Chelebaeva includes amentiferous forms comparable with recent Betulaceae and Fagaceae, which could hardly appear in the Campanian (personal communication of M.A. Akhmet'ev).

The new date obtained for the Baraba Formation substantially differs from the late Campanian– early Maastrichtian age inferred based on floral remains (Shapiro *et al.*, 1986). Formerly, the Baraba Formation was assumed to be of the early Paleocene (Danian) (Shapiro *et al.*, 1986) or Paleocene age (Slyadnev *et al.*, 1997), although it was substantiated indirectly. The U/Pb (SHRIMP) dating of zircons from tuffs of the lower Baraba Formation indicates that its accumulation commenced in the middle Eocene (after 50 Ma). The middle Eocene age of the Baraba Formation does not allow it to be considered as a facies analogue of the Upper Cretaceous Khozgon Formation, as it was assumed by Zinkevich *et al.* (1994).

Several facts should be noted, which indirectly evidence for the Eocene age of the Baraba Formation and are inconsistent with the late Campanian–early Maastrichtian age inferred from fossil flora. First, V.S. Vishnevskaya determined radiolarians of the Campanian–Maastrichtian age from tuffaceous siliciliths of the Irunei Formation overlain by the Baraba conglomerates at the Khimka River head (Shapiro *et al.*, 1986). The Irunei Formation accumulated in submarine deep settings, whereas the Baraba conglomerates are of terrestrial type. It is difficult to imagine such a rapid change in sedimentation settings during a short time interval. The Baraba conglomerates contain pebbles of various metamorphic rocks: gneisses of the Kolpakova Group, schists from the Shikhta Formation and Malka Group. There are new dates recently obtained for metamorphic rock from the Sredinnyi Range. For example, it was shown that Kolpakova gneisses were metamorphosed during two stages 77 and 47–53 Ma ago (Bindeman *et al.*, 2002). Second, the U/Pb (SHRIMP) dates of zircons indicate probably the Paleocene age of the Shikhta Formation protoliths (Hourigan *et al.*, 2001). Third, the assumed Cretaceous age is inconsistent with general appearance of the Baraba Formation conglomerates and their pebbles of acidic volcanics. All the unmetamorphosed Cretaceous sediments known in the Sredinnyi Range (Khozgon and Irunei formations) accumulated in deep-water settings: the Khozgon Formation at the foot of the continental slope (Shapiro *et al.*, 1986; Zinkevich *et al.*, 1994) and the lower layers of the Irunei Formation, in the marginal-sea basin (Zinkevich *et al.*, 1994; Konstantinovskaya, 2002). The Baraba conglomerates containing pebbles of acid volcanics were deposited in subcontinental settings. The Cretaceous acid volcanics have never been described in the Kamchatka region, and all Cretaceous sedimentary complexes are lacking any features suggesting influx of volcaniclastic acid material. In Kamchatka, acid and intermediate volcanics are known from the Paleocene Cherepanova Formation (Slyadnev *et al.*, 1997) and Eocene sequences of the Mt. Chernaya area (Gladenkov *et al.*, 1997), and pebbles enclosed in the Baraba conglomerates were probably derived form these rocks. Thus, many known facts are inconsistent with the late Campanian–early Maastrichtian age of the Baraba Formation, which is estimated based on floral data.

The analyzed relationships between rock complexes mentioned above and new data on the Baraba Formation age imply some geodynamic consequences. The Baraba Formation overlies unconformably schists and

The U/Pb age of zircon grains from dacite tuffs sampled from the Baraba Formation basal layers, Sredinnyi Range of Kamchatka

Grain no.	$\%^{206}Pb_c$	U, ppm	Th, ppm	Th/U	$^{206}Pb^*$, ppm	Uncorrected $^{238}U/^{206}Pb$	Uncorrected $^{207}Pb/^{206}Pb$	Age, Ma (± 1σ)
JG11-1	0.00	427	199	0.48	2.82	130.4 ± 3.1	0.0443 ± 4.1	49.4 ± 1.5
JG11-2	0.00	497	277	0.58	3.34	128.0 ± 3.1	0.0455 ± 3.8	50.3 ± 1.5
JG11-3	0.18	210	59	0.29	1.37	131.7 ± 3.4	0.0484 ± 7.0	48.7 ± 1.7
JG11-4	0.05	290	118	0.42	1.94	128.3 ± 3.3	0.0474 ± 5.4	50.0 ± 1.7
JG11-5	0.00	146	52	0.37	0.968	129.6 ± 3.4	0.0445 ± 7.7	49.7 ± 1.7
JG11-6	0.49	110	32	0.31	0.769	122.6 ± 3.5	0.0510 ± 7.1	52.1 ± 1.8
JG11-7A	0.07	937	107	0.12	6.81	118.2 ± 3.0	0.0477 ± 2.5	54.3 ± 1.6
JG11-8	0.00	377	87	0.24	2.40	134.9 ± 3.1	0.0469 ± 4.5	47.6 ± 1.5
JG11-9	0.00	647	283	0.45	4.40	126.5 ± 3.0	0.0457 ± 3.2	50.8 ± 1.5
JG11-10	0.14	394	105	0.27	2.67	126.8 ± 3.1	0.0482 ± 4.2	50.6 ± 1.6
JG11-11	0.43	197	97	0.51	1.35	126.1 ± 3.3	0.0505 ± 5.7	50.7 ± 1.7
JG11-12A	2.67	126	48	0.39	2.81	38.5 ± 3.1	0.0705 ± 3.7	##161.0 ± 5.0
JG11-13	0.00	790	163	0.21	5.38	126.1 ± 3.0	0.0469 ± 2.9	50.9 ± 1.5
JG11-14	0.00	327	93	0.29	2.23	126.2 ± 3.2	0.0422 ± 4.7	51.2 ± 1.6
JG11-15	0.14	391	112	0.30	2.65	127.1 ± 3.1	0.0481 ± 4.1	50.5 ± 1.6
JG11-16	0.44	333	93	0.29	2.29	124.9 ± 3.1	0.0505 ± 4.5	51.2 ± 1.6
Average age 50.5 ± 1.2 (95%) Ma, MSWO = 0.90, n = 15/16								

Note: The measurement accuracy is 1σ; (Pb_l) common lead; (Pb*) radiogenic lead. The error of standard calibration corresponded to 0.90%. Age values are calculated based on the weighted average ^{207}Pb value corrected for $^{206}Pb/^{238}U$ ages and for common Pb with an assumption that the $^{206}Pb/^{238}U$–$^{207}Pb/^{235}U$ age is concordant; (##) discordant result excluded from calculations of the weighted average age.

phyllites. Thus, it can be assumed that rocks of the Malka Group were metamorphosed prior to the middle Eocene. The Baraba rocks unconformably overly also the unmetamorphosed Irunei Formation (Shapiro *et al.*, 1986). Accordingly, it is possible to assume the pre-middle Eocene stage of marginal-sea sediments thrusting over metamorphites of the Malka Group. It should be noted also that the neoautochthon west of Mt. Baraba is overridden by the Khozgon Formation (Shapiro *et al.*, 1986; Zinkevich *et al.*, 1994; Slyadnev *et al.*, 1997). Thus, the accumulation period of the Baraba Formation was followed by additional deformation stage, when the Khozgon Formation sediments were thrust over it (Fig. 2) (Shapiro *et al.*, 1986).

CONCLUSION

(1) The crystallization age of zircons from the basal tuff layer of the Baraba Formation is 50.5 ± 1.2 Ma as determined by the U/Pb (SHRIMP) method. The accumulation of the Baraba Formation basal sediments commenced in the terminal early Eocene.

(2) The Baraba Formation overlies unconformably schists and phyllites, and, consequently, rocks of the Malka Group were metamorphosed before the middle Eocene.

(3) The Baraba Formation overlies unconformably the Irunei marginal-sea sediments thrust over metamorphic complexes, and this thrust event took place therefore in the pre-middle Eocene time.

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On the Galactic Version of Geochronological Scale

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Abstract—The galactic version of geochronological scale, which is based on hypothesis that its basic reference points correspond to moments of geological time, when our Solar System traversed the spiral arms of Galaxy, is analyzed. Concrete mechanisms of related astronomic model are discussed. As is shown, the suggested model contradicts the current understanding of Galaxy structure and the Solar System traveling through the galactic disk. It does not improve geochronological scale but leads to the inadmissible confusion.

Key words: geochronological scale, Galaxy, Sun galactic orbit, cosmic impacts.

INTRODUCTION

In their work entitled "Geochronological Scales and the Astronomic Time (State of the Art)," Barenbaum *et al.* (2002) considered the very important problem of improving the available geochronological scales to get a new one of planetary significance. They stated that a geological time scale "built with proper allowance for distribution of gravitational potential in the Galaxy and comparable in accuracy with the Phanerozoic geochronological scale, is proposed" (p. 103).

In conclusion, they mentioned, "that the problems touched" in their work "need to be analyzed and discussed much more comprehensively" (p. 112). Indeed, there are many points to be discussed. Even a brief reviewing of that work brings forth that one of the authors at least has a peculiar attitude to scientific activity. To get a deeper insight into the problem, I examined some other publications (Barenbaum, 1995, 1998; Barenbaum and Yasamanov, 1995, 1999; Pavlov *et al.*, 1991). Besides the other works, these are sufficient to give an idea of the concept under development and of the methods in use.

It is a long-known thought that the Earth is under various cosmic impacts. The problem has been considered in different aspects. A corresponding objective overview can be found in a brochure by Vladimirskii and Kislovskii (1986) with foreword written by academician A.A. Boyarchuk. There is nothing unusual in idea to consider influence of cosmic objects on evolution of the Earth and its biosphere, to correlate the cosmic and geological events. However, the problem should be considered therewith in a strict coordination with our astronomic knowledge and established facts. Like in any other scientific branch, any misrepresentation is inadmissible in this case. In fact, there are certain data suggesting three time scales for evolution of the Earth and Solar System. Explanations to this are also available (Marochnik and Suchkov, 1984). Let us consider them later on.

In this analysis, I refer to the classical book of Parenago (1954) and to capital monograph of Marochnik and Suchkov (1984), where authors considered over 800 of original works.

In essence, Barenbaum and co-authors (Pavlov *et al.*, 1991; and subsequent works) tend to convince readers that all critical events in Phanerozoic geological history are connected with epochs, when the Earth and Solar System rotating around the galactic center traveled periodically though the "galactic jet streams." As is announced, "the situation radically changed, when the phenomenon of dust-gaseous streams from the Galaxy center was established" (Barenbaum, 1998). In the same work, one can read: "Based on this discovery, it was shown" (reference to couple of works by Barenbaum) "that all important Phanerozoic events, which mark boundaries between periods of geochronological scale, are the consequences of Solar System periodic crossing of the galactic streams with regular periods of about 20–37 m.y. During those epochs ≈1–5 m.y. long, when the Sun orbit was inside the streams, the Earth could be bombarded by ~10^2–10^3 of massive galactic comets" (p. 15). In the later work (Barenbaum *et al.*, 2002), authors suggest that "up to several hundreds of thousands of galactic comets in total may fall on the Earth," i.e., >10^5 of comets. What is correct? Who did assessments? – no answer. This is a common rule. With respect to all problems concerning calculations of the Solar System's orbit or the structure of Galaxy and critical events used "to construct the scale," references to works written by astronomers are missed; the works by Barenbaum are cited only. In this connection, I feel it necessary to discuss explicitly what has been actually done by Barenbaum and his colleagues in the sphere of galactic astronomy.

PRINCIPLES OF MODELING

Principles of the concept are formulated in sections written by Barenbaum for the collective work by Pavlov *et al.*, (1991). Let us consider them in detail.

In section 22.2.1. entitled "Elaboration of Cyclicity Model," there is an attempt to define parameters of the Sun galactic orbit using formulae:

$$a = P/(1-e^2);$$

$$b = P/\sqrt{1-e^2} = a\sqrt{1-e^2},$$

where a and b are the greater and lesser semi-axes, respectively, and e is eccentricity of elliptical orbit. $P = Z/(GM_cM_0^2)$ means the ellipticity parameter, Z in which is moment of the Sun inertia relative to the Galaxy center (the original symbol is L used to denote length dimension, so I replaced it by Z), M_c is the lumped mass of Galaxy, and M_0 unexplained in the book is likely the Sun mass.

Besides many other things, expressions for a and b mean that P has the length dimension, i.e., $[P] = \mathrm{L}$ (let us use the SI system). If so, then we have a problem: according to $P = Z/(GM_cM_0^2)$, P dimension is $[P] = \frac{T^2}{LM}$ (T is time, L is length, M is mass). A discrepancy is obvious!

In the next equation (22.6)

$$R(\varphi) = \frac{P}{1+e\cos\varphi}.$$

$R(\varphi)$ is distance between the Sun and Galaxy center, and φ is phase according the author's definition. As is also evident from this equation, P must have the length dimension. Sure, the ellipsis equation in polar coordinates (Milankovitch, 1938) is

$$R(\varphi) = \frac{b^2/a}{1+e\cos\varphi},$$

where "phase" is true anomaly in terms of astronomy. Thus, it is unclear as well what the equation (22.6) with term P means.

The subsequent formula (22.7)

$$V(\varphi) = \frac{Z}{P}\sqrt{1+Ze\cos\varphi+e^2}$$

is senseless, because in addition to dimensionless parameters the radicand includes the dimensional Z. Besides, dimensionality $[Z/P] = \frac{M^2L^3}{T^2}$ has nothing in common with velocity dimension (left member of equation).

Formula (22.8)

$$t - t_0 = \int_{\varphi_0}^{\varphi} \frac{d\varphi}{(1+e\cos\varphi)^2}$$

has a continuation, but there is no sense to write it, because the left member denotes time, while the right one is expressed in radians!

In formula (22.9)

$$T_0 = 2\pi\sqrt{a^3/(GM_c)},$$

T_0 is the Sun orbital period. The expression is correct, known from reference books. It was obtained for circulatory movement $\omega = \frac{2\pi}{T}$ by means of setting equal the centrifugal (ω^2R) and centripetal (GM_c/R^2) accelerations. In the case of elliptical movement, radius R should be replaced by the greater half-axis a (the mean distance from moving object to gravity center). In the book under consideration (Pavlov *et al.*, 1991) is written: "Using formula (22.9), we define mass of Galaxy $M_c = 1.48 \times 10^{11}M_0$ that is inside the solar orbit. The calculated value is consistent with other known assessments equal to $(1.0–1.4) \times 10^{11}M_0$." The impression is that formula and former value are obtained by author.

Immediately after formula (22.9), there is expression meant to describe the mean velocity of orbital movement

$$\bar{V} = \frac{1}{2\pi}\int_0^{2\pi} V(\varphi)d\varphi = \sqrt{GM_c/a^3}.$$

The expression is incorrect, because square root from radicand does not correspond to velocity dimension. To get the correct expression from the same theorem of accelerations equality, one should use equation $\frac{V^2}{R} = \omega^2R$ (Suchkov, 1988; p. 32), and I apologize for this triviality. The correct formula is

$$\bar{V} = \sqrt{GM_c/a}.$$

Trying to figure out parameters of the Sun galactic orbit after this, author (see in Pavlov *et al.*, 1991) poses the known expressions for maximal (perigalaxy) and minimal (apogalaxy) velocities of elliptical movement (22.11)

$$V_{MAX} = \bar{V}\sqrt{(1+e)/(1-e)}$$

and

$$V_{MIN} = \bar{V}\sqrt{(1-e)/(1+e)},$$

and refers to formula (2.6) in the same book:

$$R_K(\varphi) = R_0e^{\alpha[\varphi+(2K-1)(\pi/4)]},$$

where R_K is radius of spiral K, R_0 is radius of central ring, and α is parameter characterizing the spiral coiling. He considers (2.6) as equation of logarithmic spiral applicable to galactic arms.

Then (Pavlov *et al.*, 1991; pp. 447–448) there are considerations: "... the system of equations (2.6) and (22.11) is strict, interconnecting the observed and unknown characteristics of the Sun galactic orbit." I failed to understand anything here. "Let us use equations (2.6) and (22.11) to figure out the desirable parameters a, e, and φ of solar orbit based on those orbital elements, which can be experimentally defined." Choosing distance between the Sun and Galaxy center, and the tangential component of Sun velocity relative to Galaxy center $\bar{V} = V(\varphi)\cos\beta = 233$ km/s (attention, $\bar{V}$ denoted earlier the mean rate of Sun and β is unknown angle taken from nowhere), author writes: "Finally, let us choose one of two possible solutions, in which the velocity vector $\bar{V}(\varphi)$ is consistent with direction toward the apex." Then we have an important comment: "For the Sun, apex is a point in celestial sphere, the Sun is moving toward which; it is in the Hercules constellation." I should clarify something here. Barenbaum considers always the Sun as a solitary star, and the orbit is constructed for such a star. Meanwhile, it is known well that intrinsic of Sun is also the "peculiar" movement relative to centroid of the close stars group. To be short, I give below several citations from the book by Kulikovskii (1978, p. 66). "In addition to the 'standard' (motion relative to close and bright stars of the 5–6th magnitude), there is also the so-called basic movement of the Sun relative to centroid of close stars in the Main sequence. It has the following parameters...: $V_\odot$ = 15.5 km/s; L = 45°, B = +24°." Here $V_\odot$ is the Sun velocity relative to the chosen centroid, and galactic coordinates of apex are L (longitude in the plane of galactic equator) and latitude B (angular distance from the galactic plane). This point is in the Hercules constellation exactly. Let us see further: "Finally, we obtain the Sun motion relative to the galactic center of inertia ... adding to the basic Sun movement the angular velocity of Galaxy rotation at the distance of Sun R_0 from galactic center. By hypothesis, centroid of close stars moves with this velocity $V_\odot$ = 250 km/s that is oriented along the line l = 90° (galactic longitude) and b = 0 (latitude)." Thus, the centroid movement in galactic plane is orthogonal to direction toward the center (l = 90°). Now it is clear that Barenbaum inclined the orbital plane relative to plane of Galaxy to the angle of 24° when oriented the velocity vector and, therefore, the position of solar orbit toward the apex of its local motion relative to close stars. As a result, the Sun will be leaving the galactic plane to ±24°, being able to cross the plane in orbital nodes only.

However, Barenbaum is a lucky person, getting everything he wants. In the same book (Pavlov *et al.*, 1991; p. 448), we read: "As the Sun is an ordinary star of galactic plane, let us assume value $\bar{V}$ equal to the mean rotation velocity of neighboring stars..." Further, "... the Sun orbital motion in plane Z = 0 is characterized solely by period T_0 = 250 m.y." As Z is a linear coordinate orthogonal to the galactic plane, it follows from both citations that orienting the Sun orbital plane toward the apex of its local motion, Barenbaum retains the orbit in galactic plane nevertheless! By what means I would like to know. He obtains parameters a of the greater half-axis and eccentricity e without explanations: "given these conditions, we obtain φ = 251°, e = 0.36, a = 10.2 kpc, and β = 20.1°." What is φ? All formulae in the book (Pavlov *et al.*, 1991) imply that φ is "phase" angle in elliptical motion, but formula (2.6), where φ is angle in spiral, is an exception. What is correct? What does it mean, if φ = 251°, i.e., a constant? In both formulae (2.6) and (22.11) φ must be ever varying, but there is no comments in the book. This angle cannot be found in all the figures published by authors. As for the solar orbit, it is repeated many times that it is in galactic plane. For instance, Barenbaum and Yasamanov (1995; p. 650) state: "After formation, the Sun got its own orbit in galactic plane, where it moves even nowadays..."

Putting the reality aside, Barenbaum formally states: "Moving along that orbit, the Sun is in perigalaxy at the distance of 6.5 kpc from Galaxy center and appears as far away as 13.9 kpc in apogalaxy. The Sun orbital velocity changes therewith from 364 to 172 km/s" (Pavlov *et al.*, 1991; p. 448). It is clear that Barenbaum has no idea of consequences from his construction. Let us have a look at the facts. The experimental distribution of Galaxy rotation velocities to the distance of 12–14 kpc from its center is known (e.g., Marochnik and Suchkov, 1984; p. 171). The maximum rotation velocity of central bulge (distance of 0.6–0.7 kpc from center) is up to 260 km/s according to measurements. For all the other objects rotating around the Galaxy center, velocities are lower. For instance, Marochnik and Suchkov (1984, p. 235) believe: "If the bulk mass of Galaxy would be concentrated in the central part of system, stars of galactic disk were moving along the Keplerian orbits." They state further that admissible for the Galaxy is something intermediary between the Keplerian and solid-state rotation, and that Keplerian orbits as such are unacceptable. Discussing this problem, Parenago (1954) wrote: "If there is certain gravity field, there should be out of maximum (parabolic) velocity in each point of Galaxy. Stars with faster velocities must be out of the Galaxy long ago. In the Sun neighborhood, the velocity of 300 km/s is obviously close to the parabolic one" (pp. 148–149), and this is not a unique conclusion. According to famous American experts (Bock and Bock, 1959), "velocity over 300 km/s relative to the galactic center is already fast enough to let a star to escape our Galaxy" (p. 128).

Choosing a model of solar orbit, Barenbaum (see in Pavlov *et al.*, 1991) speculated: "Such a model formally

corresponds to the case, when the radial force applied to the Sun decreases as the inverse square of distance to galactic center, and its value is determined completely by the gravity center of Galaxy $M_c(R)$... With these assumptions, the Sun motion in Galaxy must obey the Keplerian law, i.e., it must have the elliptical orbit..."

Formally, it might be so, but in reality not. It is known (Bock and Bock, 1959; p. 137) that "Orbital velocities of most stars from population I differ by 20 km/s, no more, from the purely circular velocity, i.e., by less than 10%. Their orbits are quite close therefore to the circular ones, and these stars likely reside in sites as distant from the Galaxy center as they were, when the stars were born about 5 billion years ago." The same is stated in the book by Marochnik and Suchkov (1984): "stars move in galactic disk along almost circular orbits, because the stars' peculiar velocities are one order of magnitude slower than the linear speed of their rotation. The irrotational component of their centroid velocity, which is stipulated, for instance, by the spiral structure, is of the same order" (p. 235). Similar inferences are known from works by Parenago (1954) and Kulikovskii (1978).

I should repeat that having no possibility to move along elliptical orbits with hypertrophied eccentricity parameter like $e = 0.36$ the real stars of galactic disk rotate around the Galaxy center with velocities slower than 260 km/s.

Thus, the velocity of 364 km/s suggested by Barenbaum for the Sun is much faster than the escape velocity of any star residing anywhere in the Galaxy. In situation enforced by Barenbaum upon the Galaxy, the Sun would be unable to make even a single turn. In principle, it is said enough to stop analysis of the "model." Its details are so absurd however that I cannot pass over them.

THE MODEL AND REALITY

In his paper, Barenbaum (1995, p. 30) wrote: "Structure of our Galaxy includes two individual systems of spiral arms: one with logarithmic and the other one with Archimedean spiraling. The system of logarithmic spiral arms is bounded to influence of galactic electromagnetic field and follows its configuration." Only author knows where he found these two systems; references are missed as usually. There were attempts to explain the spiral structure by influence of *magnetic field* (attention, not electromagnetic). They practically trailed off into futilities. For short, let me cite Marochnik and Suchkov (1984, p. 22): "In fact, the real situation in Galaxy is much more complicated, as is shown above, because of existence of genetically different subsystems, physically different populations of stars, interstellar gas and dust, cosmic rays, electromagnetic radiation, and magnetic fields. At the same time, an average energy density of galactic gravity field is two–three orders of magnitude greater than energy densities of other kinds... main dynamic processes in Galaxy are nevertheless under control of self-gravitating field..." Apparently, the determining factor is obvious.

In two papers published in the same year, one can read the following. "By the Galaxy rotation, the spiral construction rotates as a whole around central axis in the plane of drawing" (Barenbaum, 1995; p. 30), or "Processes of gas condensation and stars formation greatly intensify, when the dust-gaseous matter crosses four spiral arms of Galaxy, which are of electromagnetic origin (again!) and experience the differential rotation around the system center in galactic plane" (Barenbaum and Yasamanov, 1995; p. 650). What is correct then? If "as a whole," then the rotation is of the solid-state kind. If "differential," then the angular velocity should be different at points unequally distant from the rotation center. Do not wait for answer.

In the first paper is stated: "It is assumed that central mass of Galaxy, being bounded by average radius of solar orbit, did not change with time, and that solar orbit is devoid of occasional perturbations" (Barenbaum, 1995; p. 31). In the second one, we read: "... depicts one more class of rare phenomena caused by stars flyby close to the Sun that changed its galactic orbit" (Barenbaum and Yasamanov, 1995; p. 652). What is correct? Changed the orbit or not? Moreover, we read further in the same page 652: "Worthy a note in Fig. 2 is the other important trend of expotential increase of the Sun orbital period with time. A constant of this process, which is found to be equal to 5.95×10^{-11} year^{-1} (follows reference to work by Barenbaum), is explained as the result of mass loss by galactic center during evolution with the rate of 8.8 masses of the Sun per year." It is impossible to understand what is "expotential increase," but this is not a misprint, because the term is repeated in the next paragraph. Moreover, what about the central mass, did it change or not? There is no answer again.

In the other paper (Barenbaum *et al.*, 2002), we find that there were "... four significant events ..., which can be interpreted (Barenbaum and Yasamanov, 1995) as consequences of the Sun flight through dense star clouds. The interactions between the Sun and stars caused "spasmodic" changes in the solar orbit parameters" (p. 105). Thus, the orbit was not in a steady state. In the collective work (Pavlov *et al.*, 1991), Barenbaum made an attempt to estimate the influence of a close star flyby on the Sun. His results are inconsistent with assessments of professionals (e.g., Bock and Bock, 1959). In addition, nothing was said of possible close flyby rendezvous of stars in Galaxy. Meanwhile, all assessments give identical results. According to Parenago (1954) for instance, for any pair of galactic stars the average time span between events of their closest flyby is equal to 2.7×10^{15} years (p. 367). It is clear that closest flyby of stars turning their movement vector through 90° may happen once in a period many orders of magnitude longer than the lifetime of Galaxy. Even

convergence event to the distance of Neptune's orbit is estimated as one per 10000 galactic years (Bock and Bock, 1959). Meanwhile, Barenbaum and co-authors considered slightly over two last turns of Solar System and got the recurrent events of flyby through dense stellar (!) nebulae etc.

I cannot leave aside as well the idea of comet and asteroid bombardment of the Earth. An objective assessment of phenomenon is, for instance, in the book by Vladimirskii and Kislovskii (1986). In contrast, in works under consideration everything has been done to adjust the idea to geological and biotic events on the Earth. A scientific level of information retrieval by authors can be illustrated by several examples. Let us read the following in work by Barenbaum (1995, p. 34): "When the Sun travels through flows of galactic substance, very important are impacts between comets and bodies of asteroid ring. The impacts give rise to multiple generations of abundant secondary bodies and to their ejection from the ring into circumsolar space. In bulk, these bodies go away forever from the Solar System leaving in it a negligible part only, ~10^{-5} (!). Long- and short-period comets observable at present are the secondary bodies of that kind, which originated 0.6–3.6 Ma ago and were captured by the Sun, when it traveled for the last time across the Orion–Swan arm." Nobody knows how and who obtained all these values, as references are missed.

Now we can evaluate a style of scientific considerations in work by Barenbaum *et al.* (2002). For instance: "In the periods between the Sun interactions with stars, the most powerful cosmic impacts on the Earth were produced by galactic comet showers... when the Sun crossed the galactic jet streams at 19–37 Ma intervals during the Phanerozoic time" (p. 106). The estimated frequency of such events was cited above, and the issue will be discussed one more time below based on the model by Barenbaum. Let us go further: "Nearly 200 events of the Phanerozoic time can be interpreted (!) ... as the impact results of solitary asteroids more than 3.5 km in diameter." Yes, one can have a chance to interpret, but how it was in reality? After all, the purpose is to improve the time scale. Next: "In the intervals of several million years, when the Solar System occurred inside the galactic comet stream, one or several big asteroids might or might not fall (!) on the Earth" (p. 108). Everything could happen, but what is the sense of model then? "In the most ME epochs, at least one big asteroid fell on the Earth during the comet bombardment" (p. 108). It fell for all that! However, where is proof not "ex contrario," not inferable from mass extinctions? It is necessary to establish independently the cause of extinctions, but not vise versa. "As a rule however, asteroids fell after initiation of the comet bombardment" (p. 108). What is evidence for such categorical statements? We let everybody to judge the statements personally. What is strange however, devastating arguments against all the speculations can be found in the Barenbaum's "model" as it is. Let us consider two figures only and related author's comments.

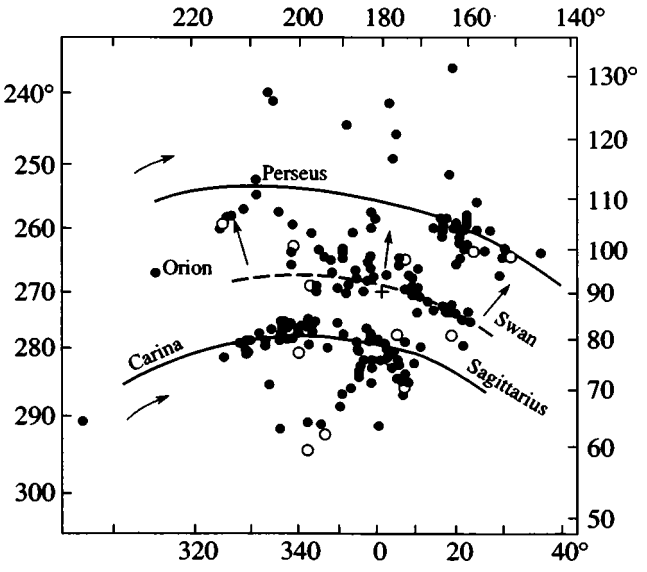

Fig. 1. Clusters of young stars (black dots) and stellar O-B associations (open circles) near the Solar system (+) projected on the galactic plane; solid and dotted lines denote logarithmic and Archimedean spirals; substance flow directions in spiral arms are shown by arrows (reproduced Fig. 2 and caption from Barenbaum, 1995).

In the book (Pavlov *et al.*, 1991; Fig. 22.2) and in a paper (Barenbaum, 1995; Fig. 2), there is a figure reproduced from an earlier publication (Bock and Bock; 1959 Fig. 120). Published later on in a final form (Bock and Bock, 1981), the figure was also reproduced by Marochnik and Suchkov (1984, Fig. 109) with necessary references, which are missed in the aforementioned book and paper, where authors perverted the original and changed number of stars and other objects (see Fig. 1). Their creative work resulted in appearance of logarithmic (Perseus, Carina–Sagittarius) and Archimedean (Orion–Swan) spiral arms. Here I should give a citation (Barenbaum, 1995; p. 31): "This arm differing (Marochnik and Suchkov, 1984) from the Perseus and Carina–Sagittarius arms by the other pitch angle and kinematics of stars is nothing else (Barenbaum, 1993) but a front of jet stream." References inside citation are given by Barenbaum. Now I cite the original (Marochnik and Suchkov, 1984; p. 340): "Here we see (as far away as 3–5 kpc) two major elements of the global two-armed structure of Galaxy: the Sagittarius–Carina arm ("internal" arm tending toward the Galaxy center) and the Perseus arm ("external" arm at the distance of 12–13 kpc away from the center if R = 10 kpc). In addition, optic indicators detect one more clustering of young stars and nebulae, the Orion arm with the Sun inside. At present, there is almost no doubt that the Orion arm is a cross-connection between two major arms rather than an independent element of the global galactic structure." Thus, it is not a "front of jet

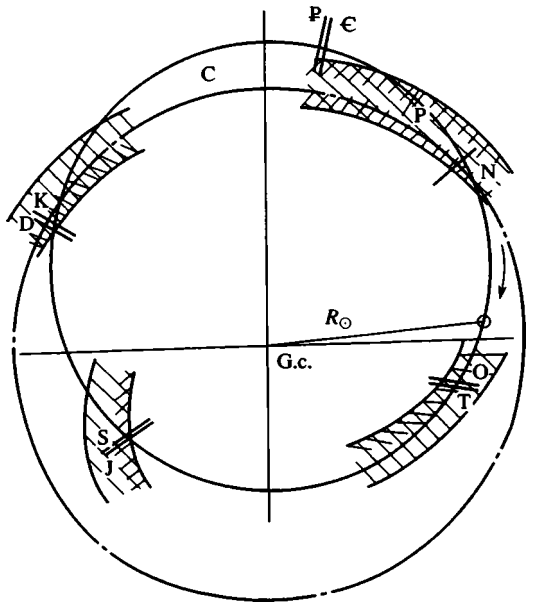

Fig. 2. Solar orbit projected on galactic plane: (G.c.) galactic center; ($R_\odot$ = 10 kpc) current distance between the Sun and galactic center; arrow shows direction of the Sun orbital movement. Letter symbols near line segments intersecting the orbit denote boundaries between periods of Phanerozoic geochronological scale. Galactic arms traversable by the Sun are hatched; the Perseus and Carina–Sagittarius arms are in the upper and lower halves of the figure, respectively (reproduced Fig. 1 and caption from Barenbaum and Yasamanov, 1995).

steam," Barenbaum filled with which the Galaxy. Next, in opinion of Barenbaum the are four logarithmic and two Archimedean spirals seeming to be invented by him, because references to observations or to conclusions of specialists are absent. In fact, specialists in majority are adherents of the two-armed model but not of a "mixture" created by Barenbaum.

In the papers (Barenbaum, 1995; Barenbaum and Yasamanov, 1995, 1999), there is also a figure (see Fig. 2 in this work) meant to explain how the Sun periodically crossed the galactic jet streams. It is interesting to compare Figs. 1 and 2. It is clear from explanations to Fig. 2 that the Perseus and Carina–Sagittarius arms also shown in Fig. 1 are placed respectively into upper and lower halves of Fig. 2. However, where is the Orion arm with the Sun inside in the latter figure? Instead, authors set the Galaxy center between the former two arms and launched the Sun to travel along an elliptical orbit crossing those arms. I cannot understand how it was possible to set the Carina–Sagittarius arm behind the center (observers detected it at the distance of about 2 kpc toward the Galaxy center)! There is more to come yet. According to Barenbaum and Yasamanov (1995): "The Sun orbital period was 250 m.y., i.e., almost identical to rotation period of galactic arms. Accordingly, space areas of galactic arms and jet streams, which are crossed by the Sun during the first and second turn, fall, as is shown in Fig. 1 (Fig. 2 in this work), in the same segments of solar orbit" (pp. 651–652). It is impossible however. If the rotation periods of the Sun and galactic arms are identical, the Sun will be motionless in average relative to that point in the arm, where it is at the given moment (after all, they rotate simultaneously). If the Sun moved nevertheless, as is shown by arrow, and crossed the arms, then it rotated with that period of 250 m.y. relative the arms! Did it move two times faster? If so, what should we do with the "escape" velocity?

Absurd?! Of course. Let us take, however, the identical rotation periods. They obviously exclude for the Sun moving around the Galaxy center a possibility to cross successively the spiral arms. What we have in the rest? As the solar orbit is elliptical, we have instead the Sun epicyclic movement relative to a certain mean place that uniformly moves together with galactic arms with the same period of 250 m.y. Such a situation known well in astronomy resembles "center equation" entering the equation of time. There is the same. In distinction from the uniformly moving mean Sun, the real ecliptic Sun moves nonuniformly, and both pass simultaneously the terminal points of the apse line. During one half-period, the real Sun first gets ahead of the mean Sun, while the latter gets up to the former by the half-period end. In the next half-period, they travel in reverse order. In the Barenbaum's construction, the first half-periods begins, when the Sun orbital velocity is at maximum after passing the perigalaxy.

Leaving aside the details (they are described in manuals of astronomy), I should repeat that the Sun will be in epicyclic movement around a centroid, given the equal periods of its orbital movement and the galactic arms uniform rotation. The theory of epicyclic movement is known (Marochnik and Suchkov, 1984), and the necessary path can be easily calculated. The fact is that the Sun must always be, as it actually does, in vicinity of the arm, to which it belongs, thus being unable to cross the galactic arms one after another. Of course, Barenbaum and his colleagues assumed an abnormally high eccentricity value $e = 0.36$, but this can let the Sun to reach the adjacent arm only and nothing else. As is shown however, with such an orbit the Sun will escape our Galaxy because of the extremely high velocity at the perigalaxy point.

Of course, the "newest" speculations and illustrations of Barenbaum *et al.* (2002) are of interest for analysis in order to stress their ineptitude, but this will not change anything. I leave without comments their "new" period of the Sun orbital movement (p. 110) and confusing words about "the anomalistic period of the Sun motion (250 m.y.)," which is "a multiple of the rotation period of the core disk of the Galaxy (50 m.y.)." I want to cite only the next paragraph: "With due account for distribution of the gravitation potential in the Galaxy, it was figured out that the orbital period of stars of the galactic plane, at the distance of Sun from the Galaxy

center, should be 223 m.y., i.e., slightly lesser than the anomalistic period of the Sun (250 m.y.). This circumstance implies the slow turning of the apse line of the solar orbit by $\sim 1.4 \times 10^{-7}$ angular degrees per year toward the direction of the Sun movement."

First, I would like to know what did they for the "due account for distribution of the gravitation potential in the Galaxy?" There is no answer; the statement is unsubstantiated. Second, the suggested then turning of the apse line can be easily calculated based on difference between phenomena frequencies. If the rotation periods are known, then the turning period of apse line is

$$\frac{1}{T_A} = \frac{1}{223} - \frac{1}{250},$$

all periods in m.y. Barenbaum and his colleagues may check that their own data give the rotation angle of apse line, which is $\sim 1.74 \times 10^{-7}$ but not $\sim 1.4 \times 10^{-7}$ angular degrees per year, as they state! This is nothing to speak of. The main thing is general approach with the other illustrations now.

Let us read the following (Barenbaum *et al.*, 2002; p. 111): "The arms are tentatively considered as 1 kpc wide. It is a general opinion that, simultaneously with the Sun orbital motion, the entire four-armed spiral construction (Fig. 1) rotates around the center at the angular rate $\omega_0 = 2.8 \times 10^{-8}$ rad/year that gives the rotation period of 223 m.y. The newly calculated moments, when the Sun crossed the galactic streams, are indicated by numbered points on the solar orbit."

Therefore, the Sun and "four-armed" construction have now the identical rotation periods of 223 m.y. As is already shown, the Sun would be in this case near its mean position being unable to cross the galactic arms (the case of simultaneous rotation!). Authors illustrate the solar orbit in their Fig. 1 and countdown time of the orbit relative to spiral arms in Fig. 3b. Both figures are interrelated (in Fig. 3, there is reference to Fig. 1: Roman numbers denote spiral arms of the Galaxy). I do not reproduce these figures here. Anybody who is interested may try to figure out, comparing originals, in what order the Sun crosses "spiral arms" and "jet streams," admitting that crossing took place. You may begin to count right to left following the geological time succession of abscissa (see simultaneously Fig. 3b and numbered points of orbit in Fig. 1 of Barenbaum *et al.*, 2002). No good will comes of it, a mess only. If vise versa, the result will be worse! One cannot expect anything else. The authors' statement that rotation periods of the Sun and galactic arms are equal discredits the invented construction! By no means the "model" substantiates the Sun–galactic arms intersection with periods of 20–37 m.y.

It would be better to omit all the countless assumptions. It was just cited (see above) that "arms are tentatively considered as 1 kpc wide," but in other works (Barenbaum, 1995; Barenbaum and Yasamanov, 1995) is written: "...both arms are 2 kpc wide..." Where is the truth, and where is its violation? Well, now about conclusion to paper by Barenbaum *et al.*, (2002) where is said: "The considered data allow us to assume the following..." Thus, all this is only assumed?!

CONCLUSION

It is impossible to comment here all the contradictory speculations present in the discussed papers and in the book by Pavlov *et al.* (1991). One thing is clear however. Authors claim with no ground at all: "the proposed geological time scale ... is comparable in accuracy with the Phanerozoic geochronological scale." The farfetched scheme perverting well-known facts of galactic astronomy destroys itself all speculations of authors.

Taking into account everything that was said, I cannot leave aside all the talks about "new paradigm in geology." In the conclusive paragraph 22.3 of the manual written by Pavlov *et al.* (1991), one can read: "It is possible to suggest four postulates, which outline rather completely, as we think, the new quantum paradigm in geology.

1. Geological evolution of the Earth is determined by energy received from outside.

2. The Earth receives energy quanta of about 10^{30}–10^{31} joule in a short time span (1–5 m.y.).

3. Energy quanta originate when Solar system crosses the galactic jet streams and are related to bombardment of the Earth by comets and meteorites, and by close flyby of stars responsible for transformation of solar and terrestrial orbits.

4. Periodicity of Solar system traveling through galactic jet streams determines the periodic receiving of energy quanta.

One can easily recognize here the "model" analyzed in this work. Now it is in a form of postulates, the categorical statements, which must be learned even if they contradict facts. How sad that this unsubstantiated "paradigm" is declared in the manual for higher education!

Reviewers Yu.N. Avsyuk and N.M. Chumakov

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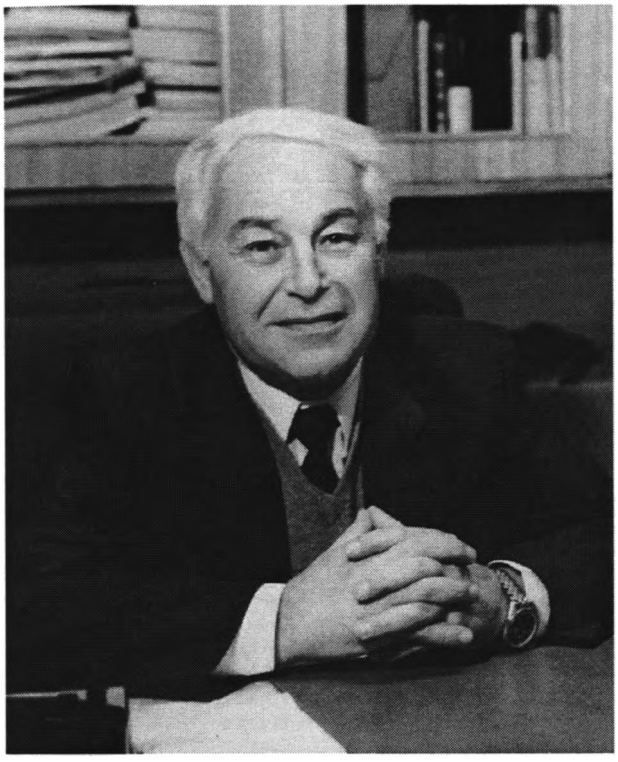
In Memory of Nikita Alekseevich Bogdanov (1931–2003)



On December 14, 2003, Nikita Alekseevich Bogdanov, Corresponding Member of the Russian Academy of Sciences (RAS), Director of the Institute of the Lithosphere of Marginal Seas (IL RAS), Professor of Moscow State University (MGU), a member of the Editorial Board of *Stratigraphy and Geological Correlation*, passed away after a short decease. Brought to an end was the life of a wonderful person, an eminent scientist, and outstanding organizer of science. Nikita Alekseevich was born on July 23, 1931, in Astrakhan to the family of a geologist, the Dean and Professor of the Geological Faculty, Moscow State University. In 1954, he graduated from the Ordzhonikidze Moscow Geological Exploration Institute and joined the Geological Institute, Russian Academy of Sciences. In 1978, he was invited to the Laboratory of Sedimentary Mineral Resources. Soon afterward, the laboratory was transformed owing to Bogdanov's organizing activity into the Institute of the Lithosphere, RAS, renamed later as the Institute of the Lithosphere of Marginal Seas. Bogdanov was permanent head of the institute for the last 15 years.

Bogdanov was an expert in geology of the ocean and marginal sea floor, known well in this country and abroad, the author and co-author of 260 scientific publications, including 11 monographs and 9 tectonic maps.

Based on studied structures and evolution marginal seas and World Ocean, Bogdanov created a concept of the development of basins floored by oceanic crust. He has generalized the most recent material on the structure of shelf seas of Russia and outlined basic trends in formation of the sedimentary cover basement in shelf zones. His scientific results are in wide use by oil-and-gas prospecting. He created a modern concept of the Arctic basin structure.

For many years, Bogdanov took charge of the UNESCO international scientific projects and befittingly represented the national science abroad. He was representative of the USSR and later Russia in the Planning and Executive committees of the Ocean Drilling Program. He found remarkable favor and respect amidst foreign geologists, was an honorary member of Geological Societies of the United States, Australia, and France. His election to the post of Secretary General of the 27th International Geological Congress that was held with a great success in Moscow in 1984 was the recognition of his merits. It was then during the preparation and performance of this important international forum that his brilliant organizational gift showed up.

As Professor of Moscow State University, he delivered a special course on plate tectonics of the World Ocean. Bogdanov supervised six DSc and fourteen PhD dissertations.

In 2003, Bogdanov and a group of other scientists were awarded the State Prize of the Russian Federation for a series of tectonic maps of Arctic seas and adjacent territories.

A striking trait of Nikita Alekseevich was his grateful attitude to teachers and predecessors N.S. Shatsky, B.M. Keller, V.V. Menner, and many others, whom he often recalled with tenderness. He was astonishingly active person, no matter whether it concerned the routine life of the institute or events taking place in the Academy of Sciences, Moscow, and the country as a whole.

The passing away of Nikita Alekseevich Bogdanov is a great loss for science, all his friends, collaborators, and disciples. He was a frank, benevolent person, and his unexpected going away was a shock for everybody. Nikita Alekseevich will forever remain in the hearts of all who associated with him at work and through life.

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